CHAPTER 6

DISCUSSION AND APPLICATION

6.1 SUITABILITY OF METHODS

6.1.1 LABORATORY TECHNIQUES

A large effort was expended in the laboratory to verify azeotropic distillation as a method to obtain water for isotopic analysis from soils, and particularly to extend that or another technique to gypseous soil. While this was motivated by the hope of applying the Barnes and Allison (1983) type convection-diffusion model to estimate discharge in the field area, extraction of water for isotopic analysis is relevant to other types of studies as well. While azeotropic distillation had been used for nearly a decade in the CSIRO laboratory, little detailed work on the method was done. Close scrutiny in this work confirmed it to be a cheap and practical method for a large range of soil types, although the accuracy was not always as good as some early quoted figures. Distillation with kerosene is fairly quick (two or three runs per day), and the combined extraction and analytical uncertainties add little or nothing to the uncertainty of the derived data, principally evaporation rates.

The uncertainty involved using hexane distillation to obtain pore water from gypseous samples is considerably larger, but because of the availability of multiple data points in each profile from which calculations were made, most of which were obtained using kerosene distillation, accuracy was still not a problem.

The errors associated with dilution extraction were however significantly large for the purpose of the data. Because of large scatter, dilution extraction data were not used in the interpretation of profiles. Vacuum distillation at room temperature proved too unreliable, and while adsorption of vapour by desiccants showed some promise, much more work would be required before it could be applied.

Other laboratory techniques were standard. The measurement of gypsum content was quite approximate, and the method suffers from the contribution of water from clays to give an overestimate of gypsum. This did not present interpretational difficulties, but did slow the progress of extraction of field samples, until it was realised that some "gypseous" samples contained no gypsum, and could be distilled with kerosene, rather than be left waiting for the trials to produce the most suitable method for gypseous soils.

The interlaboratory, international comparison of various methods to determine the stable isotope composition of three soils, one with two moisture contents, showed an alarming spread of results. The preparation of the samples was a potential source of error, but the standard deviations of replicates of each soil type used, and another considered in preliminary trials, were quite small, varying between 0.5 and 1.2‰ for δD, and 0.13 and 0.48‰ for $\delta^{18}O$. Shipping by air, with the possible freezing and thawing, or leaks aggravated by pressure changes, might possibly have induced some of the scatter; this is considered unlikely by the author and other workers in the same laboratory who have used the same containers for storage of soils for many years, but cannot be excluded as a possibility. Values obtained by azeotropic distillation tended to be isotopically heavier than those obtained by vacuum distillation at various temperatures, with some exceptions. The range of values obtained, means and standard deviations for the three non-gypseous

soils are given in table 6.1.1, together with the result by the author;

this is meant for comparison and is not a claim that my value is the most

correct.   Indeed, my results are the heaviest for two of the non-gypseous

samples.

The results for the gypseous soil, number 2, need to be considered

separately (Tab. 6.1.2).   Most methods obtained both gypsum and pore

water from the sample, with a fairly small scatter.   The best measure of

pore water composition is by that sample extracted by centrifugation, and

that obtained from hexane distillation agrees fairly well when the

correction mentioned earlier is applied.   The oxygen results do not

TABLE 6.1.1

Summary of Results from the Interlaboratory Comparison of Methods to
Determine the Stable Isotope Composition of Soil Water, Non-Gypseous
Soils.   Incomplete results to December 1989.

Soil	water content $g\ g^{-1}$	no. analyses	Range	Average $\pm$ s.d.	Author's value
			---------------- δD, ‰ SMOW -------------------		
1. Sand	0.051	11	-49.1 to -22.5	-29.0 $\pm$ 4.8	-24.6
3. Clay	0.147	9	-57.1 to -28.1	-39.9 $\pm$ 8.4	-28.1
4. Clay	0.252	10	-44.2 to -24.0	-33.6 $\pm$ 5.8	-24.0
			------------- $\delta^{18}O$, ‰ SMOW ----------------		
1. Sand	0.051	7	-5.03 to -3.81	-4.46 $\pm$ 0.37	-3.81
3. Clay	0.147	7	-8.02 to -3.61	-5.56 $\pm$ 1.29	-3.61
4. Clay	0.252	7	-6.57 to -3.43	-4.74 $\pm$ 1.05	-3.43

TABLE 6.1.2

Summary of Results from the Interlaboratory Comparison of Methods to Determine the Stable Isotope Composition of Soil Water, Gypseous Soil (No.2). Incomplete results to December 1989. Water content was 0.102 g g^{-1}.

Method	No. reps.	δD ‰	δ^{18}O ‰
Centrifugation	1	-20.0	-3.27
Azeotropic Distillation with Hexane	10	-23.2±1.2	-3.75±0.20
All Other methods	9	-30.6±3.6	-2.86±1.33
Gypsum Crystal Water	6	-35.5±1.1	+2.92±0.24

follow the same pattern as the deuterium, with some techniques obtaining close to the same value as centrifugation and hexane distillation, despite the deuterium value on the same water sample being quite different.

The overall scatter of the full data set is much greater than that claimed for any of the methods individually, and appears to be both material and water-content dependent. The intercomparison points out the need for careful laboratory technique, and the verification of technique for different soil types. Comparisons of the absolute values of soil water isotope data obtained by different laboratories should be made with caution when different techniques are used, or even the same technique at a different temperature or with a different hydrocarbon.

6.1.2 CONVECTION-DIFFUSION MODEL

6.1.2.1 Limitations of Parameters and Assumptions

Effective Diffusivity

The diffusivities of stable isotopes of water and Na Cl solutions are well known, but the adjustments to calculate effective diffusivity in porous media uncertain. In the approach adopted in this work, two modifying factors are used, the volumetric water content θ_v and impedance factor f_l. The gravimetric water content θ_g may be fairly well measured, but not without uncertainties. Field variability is usually high, and determinations replicable to about 0.005 g g^{-1}, or between about 1 and 10% for the field samples. Vacuum-dry water contents are invariably a little less than oven-dry, and it is sometimes difficult to tell when samples are truly "vacuum-dry", as weight loss is very slow in the later stages of drying. The conversion of gravimetric to volumetric water contents is subject to uncertainty of the bulk density of the soil. This was measured fairly accurately on samples of solid shale, but estimated only for alluvial sediments, soil, gypseous material and weathered clay. Another approximation is choosing an average water content for the segment of profile used in evaporation estimations; this was done visually from the water content profiles.

Another potential problem largely ignored is that of non-participating water in the excluded volume involved with anions such as chloride. This reduces the cross-sectional area available for diffusion, and hence effective diffusivity. The problem is however incorporated into the impedance factor, which is measured as a function of the total (easily measured)

water content. Allison and Barnes (1985) allowed for the partial volume of sodium and chloride ions in concentrated solution for pore water at Lake Frome, where the reservoir concentrations were high, from 98 to 184 g l^{-1}. This correction was not considered justified for this project.

The greatest uncertainty in estimating D_{eff} is in finding a realistic impedance value (c.f. Barnes & Allison, 1988). Few measurements for water isotopes are available in the literature, the best data that could be found was discussed in section 2.2.4. The parameter is clearly material dependent, and varies with water content. Over the range of water contents of the field samples, the use of a constant f_l of 0.5 was as good as any other formulation, in the absence of data from the field materials. The value may however be out by up to 40%, going by the published data.

Many measurement of f_l for chloride are available but they show a large scatter. They invariably decrease with lower water content, and usually become zero at non-zero water contents. The form adopted, $f_l = 0.8 \, \theta_v - 0.05$, is far from a unique solution, and was picked as plausible for clayey material. It is difficult to assess what the error involved is, but to say it is similar to that for f_l of deuterium is probably fair.

Some of the parameters used in calculations are temperature dependant. The temperature of soil in the field area was not measured, and a temperature of $25^\circ C$ implicitly used. This is reasonable for the latitude of the project area, and close to the temperature of the groundwater from bores where the aquifer is closer to the surface and not so affected by the geothermal gradient (23 to $26^\circ C$, A.G.C., 1984).

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Boundary Conditions

The simple convection-diffusion calculations assume both top and bottom boundary conditions to have constant concentration of the solute modelled. With chloride, this is provided by the presence of solid sodium chloride in the soil at the top boundary, and by the deep aquifer water value below. The latter (1 g l^{-1} Cl) is so small compared to the former (220 g l^{-1}) that no extra errors in evaporation estimates would be induced if the aquifer value were out by 100 or 200%, which it is not. The *use* of the deep aquifer value to provide the bottom boundary condition is a little problematical at Hamilton Hill, where the shallow water table exists in interbedded alluvial sands and clays, and other water sources (local recharge, flood water) are possible. The similarity of the isotopic signature of the shallow groundwater to that elsewhere suggest it comes from the same deep source, the evaporation estimates using the deep groundwater boundary condition are more in keeping with those from similar water table depth elsewhere, and the chloride and deuterium estimates agree more closely than with the shallow groundwater boundary condition.

The top boundary condition for chloride is only in some doubt in the flood plain holes, as halite saturation is not always present in these fresher profiles. Halite is present for much more time (between large rainfall events) than it is not, and the characteristic times are in the order of years. The top of these profiles, as with deuterium, varied considerably with seasonal effects, but were invariant below about 0.7 m. Taking this point as the top boundary condition would have little effect on the interpretation as the shape of the $\ln(c-c_{res})$ versus depth curve is used, and c_o, the concentration at the top boundary, is not involved.

It is likely that some halite from the salt lakes of the area is blown from the lake surfaces when dry, and deposited with dust on the surrounding land surface. Any such salt would then be washed into the soil profile by later rains. The presence of the mineral halite in the soil profiles ensures that the top boundary condition required for the evaporation estimates is still met. Addition of sodium chloride from above would tend to increase the quantity of halite in a soil profile, and such may be a factor in the case of hole 13, near the shore of Lake Eyre South, which evidences a thick, halite-rich layer in its profile.

For stable isotopes, the top boundary condition is maintained by the balance of heavy isotope "excess" produced in the evaporation zone (or rarely, "front") and diffusion away above, to the atmosphere, and below, to the water table. With the Hamilton Hill profiles, seasonal effects near the surface are important, and the top boundary condition is very much an "average". With most of the gibber plain holes, the isotope peak was very broad, and hence probably very stable, at depths below seasonal influences, forming a good constant concentration boundary, although the exact location is difficult to define.

All sites were chosen where vegetation was absent, as its presence invalidates the assumption that all water is lost to evaporation, as root extraction is largely non-fractionating (Barnes & Allison, 1983, and others.)

Steady-State

The other major assumption of the convection-diffusion model used is that the flux of solute and water is steady. This seems quite possible in the field area where the climate is arid and has been for a few thousand years at least. Gibber plain holes, except hole 24, were

deliberately sited at slight rises where ponding and large infiltration of rainwater is unlikely. This assertion was supported by the evidence of holes 22 and 23, drilled after the extraordinary rains of March 1989, and showed very minor, shallow infiltration, in contrast to deep infiltration and profile modification at hole 24 in a slight hollow. The effects of infiltration on most profiles did not affect the shape of the solute profiles below about 0.7 m, so that evaporation estimates from the profiles that were interpreted may be treated with some confidence.

The time scales of the profiles are in some cases very long, exceeding 10,000 years. In this time frame climate and even topography cannot be assumed to be unchanging. The profiles would be very slow to change in response to a new evaporation rate if the depth of the water table changed, due to, say, erosion or deposition of soil, increased or decreased leakage from the deep aquifer, or if the evaporative power of the atmosphere changed with a different average relative humidity. The evaporation values are very much long-term averages. The changes themselves are likely to have been slow, in the main part. The difference between a late-stage non-steady and a steady-state profile at the same evaporation rate is small (Barnes & Allison, 1988), and its interpretation as a steady-state profile does not induce a large error in estimated evaporation rate.

Lateral Movement at the Water Table

Some holes show an obvious discontinuity at the water table, notably holes 8 and 9 (section 4.4.2.1). Convection-diffusion calculations to estimate evaporation were not justified on such holes, and the discontinuity is assigned to lateral water movement within the water-table aquifer. The groundwaters from such holes are in agreement

with the general pattern of chemistry and isotope content for the area, with its major source leakage of artesian water from below. It can be expected that the actual evaporation of shallow groundwater is similar to other sites where the water table is at a similar depth. Leakage from the deep aquifer may well be greater than at the other sites, if the fractured zone of shale extends well below the water table into the aquitard.

6.2 EVAPORATION ESTIMATES

6.2.1 ADOPTED VALUES AND UNCERTAINTIES

A summary of estimates of evaporation, some rounded off, and the adopted (preferred) values are given in table 6.2.1. They are plotted against depth to water table in figure 6.2.1. The adopted uncertainty is 50%, following the considerations outlined in the previous section 6.1. The greater evaporation form holes 4 and 18, compared to 12 and 20 with comparable water tables, is probably due to the heavier texture at the former, contrasting with the sandier profiles at the latter. Sand acts more like a mulch (i.e. the vapour path length is longer because of the deeper layer of very low water content material, so that the vapour gradient and flux are reduced; see Allison *et al.*, 1983), than silt and clay. Although the gibber plain sites are clayey in the weathered Bulldog Shale, the tops of the profiles are sandy, or consist of coarse silt, and the evaporation rates are more in accord with holes 12 and 20.
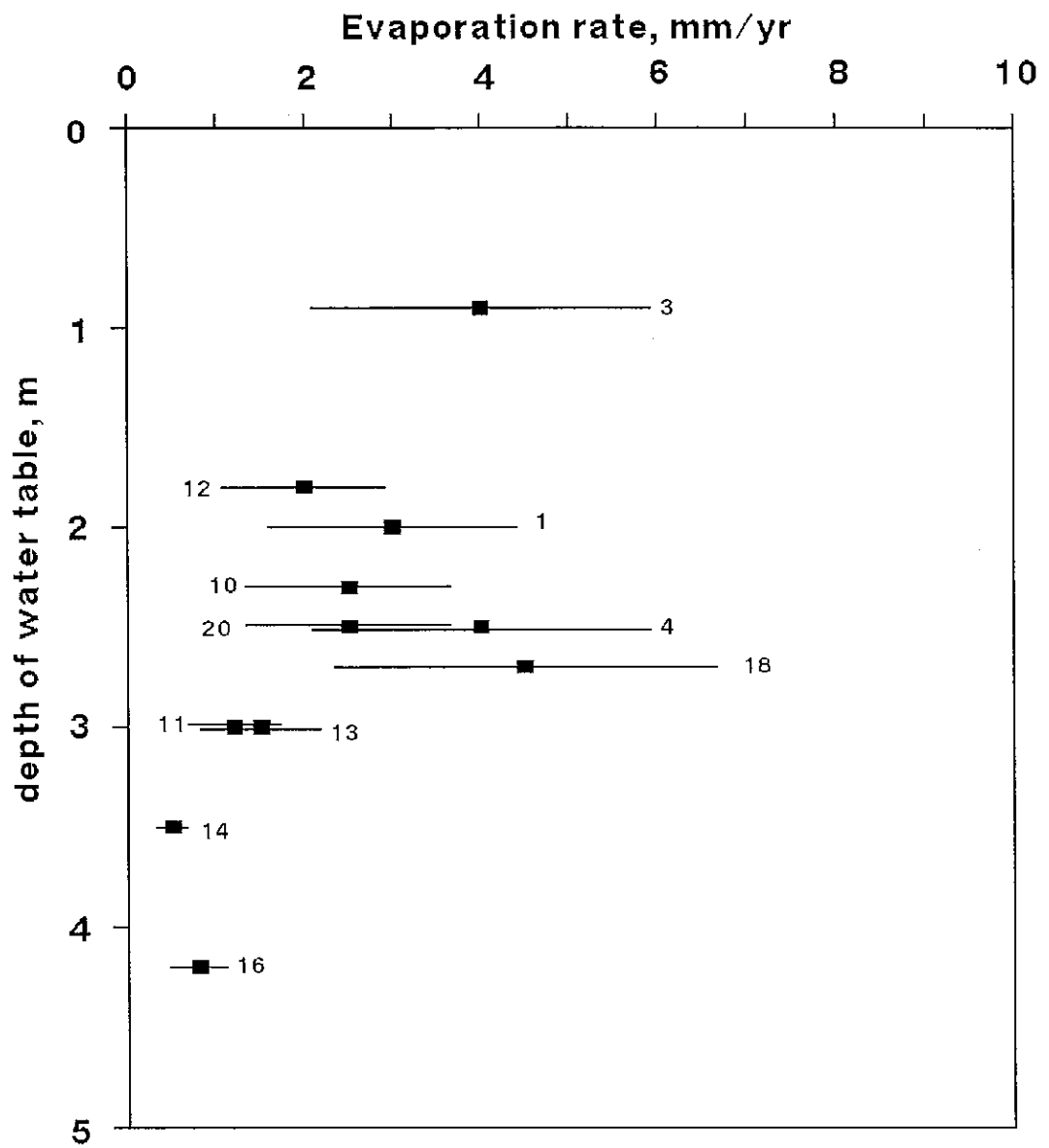
Figure 6.2.1 Evaporation vs water table depth, project data

Adopted values with estimated uncertainty (50%)

hole numbers next to data

TABLE 6.2.1

Adopted evaporation estimates

Hole	Water Table	Evaporation, mm yr^{-1}				
	m bgl	Deep b.c.		Shallow b.c.		Adopted
		Cl	δD	Cl	δD	
1	2.0	3.0	–	19	–	3.0
3	0.9	4.0	4.0			4.0
4	2.5	4.0	4.0	1.4	8	4.0
10	2.3	3.5	1.6	–	–	2.5
11	3.0	0.8	1.8	–	–	1.5
12	1.8	1.4	2.8	5	9	2.0
13	3.0	1.3	1.1	–	–	1.2
14	3.5	0.4	0.5	–	–	0.5
16	4.2	0.7	0.9	–	–	0.8
18	2.7	5.0	4.3	6.5	13	4.5
20	2.5	2.1	2.3	3.3	5.7	2.5

bgl = below ground level.   b.c. = boundary condition

6.2.2 COMPARISON WITH OTHER RESULTS

The data of this project are shown with the data from Lake Frome of
Allison and Barnes (1985) and from East Saharan depressions of Christmann
and Sonntag (1987) in figure 6.6.2. A logarithmic scale of evaporation
has been adopted to enable the range of values to be displayed. These
two sources were chosen as the closest similar published work that had
several estimates made with different depths to water table.

This project's values are on the lower end of the evaporation range
and deeper part of the water table depth range compared to Christmann and

FIGURE 6.2.2 Evaporation vs water table depth: comparison of project and other published data

✕	Christmann & Sonntag, 1987 (no errors quoted)
─▭─	This project
⊖	Allison & Barnes, 1985 (Statistical fitting error only)

Sonntag's work, and much less than the values of Allison and Barnes, which had very shallow water tables. All these other values were calculated using an f_l of 0.66 (deuterium), compared to 0.5 for the current work. Evaporation increases greatly as the water table approaches the surface as expected (c.f. Talsma's (1963) curves, Fig. 1.4.1). The points do not trace a smooth, monotonic decrease in evaporation as depth to water table increases, as should be expected from the uncertainties involved and the variation of soil type. The Lake Frome holes were in fine-grained sediments with an efflorescent salt crust, while most of the East Saharan holes were in sandy sediments, several with pronounced salt crusts.

Other evaporation estimates with much greater depth to water table are those in the West Sahara (section 1.4.3.3). These had evaporation rates of 1 mm yr^{-1} from a water table about 20 m deep (Zouari, 1983), 2 mm yr^{-1} from about 10 m (Yousfi, 1984), and 1 to 7 mm yr^{-1} from 7.2 m and below (Colin-Kaczala, 1986). These estimates are all greater than would be predicted by the trend of figure 6.2.2. All were made using $f_l = 0.66$ for deuterium, but the use of 0.5 would only reduce the numbers by 24%. The main difference in the profiles seems to be the lack of halite (or very sparse occurrence at the soil surface). Saturation with Na Cl lowers the relative humidity of the evaporation zone, reducing the gradient driving the vapour flux, and makes condensation, mainly at night, a possibility, again reducing evaporation (c.f. Kinsman, 1976). It is also noted that the Sahara is more arid than the Lake Eyre region of Australia, with average annual rainfall, pan evaporation and average relative humidity at Béni Abbès being 32 mm, 3200 mm and 38%, compared to Lake Eyre South with (about) 130 mm, 3000 mm and 40%.

The evaporation estimates in the project area are much less than those predicted by Talsma's curves, which would suggest rates of 150 to

700 mm yr^{-1} for water tables between one and three metres, using the more conservative Kerang-Wakool-Deniliquin curve (Fig. 1.4.1). The major difference is that the sites he considered are vegetated and (relatively) non-saline. Plants increase evaporation compared to a bare soil under the same climate, by the extraction of water by their roots, and may sometimes take water from the water table itself (section 1.4.2.5).

6.2.3 RELATIONSHIP BETWEEN EVAPORATION AND LEAKAGE

The convection-diffusion calculations measure the flux of water coming from the water reservoir, and in most field profiles this means leakage from the deep aquifer. If the shallow water-table constituted the boundary condition, then it is strictly evaporation from the water table that is estimated. In lowlying or fractured ground there may be an additional source of water, such as local recharge or laterally introduced groundwater. Some laterally introduced water may have been originally vertical leakage elsewhere, and there could be a nett lateral loss of water. Evaporation should be about the same as that from another site with similar soil/sub-soil material, and the same water table depth. It would give an idea of the discharge, yielding an overestimate if there is an additional nett source of water, or underestimate if there is a nett loss.

Most of those field profiles interpreted by the convection-diffusion model would appear to give a direct estimate of leakage, as the profiles needed to show little sign of lateral flow and infiltration to be analysed by that model at all.

6.3 <u>AREAL</u> <u>ESTIMATION</u> <u>OF</u> <u>LEAKAGE</u>

If the depth to water table were known throughout the project area, evaporation as an estimate of discharge could be assigned using a "smoothed" version of the evaporation versus water-table depth curve of figure 6.2.1 or .2, perhaps trying to make some allowance for the patchy recharge observed after the March 1989 flood. The data is unavailable, and unlikely to be ever gathered in detail, except perhaps in the borefield proper. The most appropriate approach to extending the available estimates is to assign leakage rates to physiographic units. Four units have been chosen, based on topography and the extent of the aquifer from geophysical surveys (Tab. 6.3.1, Fig. 6.3.1), with low and high cases proposed to try and allow for the uncertainties in the figures. The calculated averages are for diffuse discharge by leakage to the water table for the project area as a whole, and averages referred only to the areas underlain by significant aquifer material, the sub-basins.

The diffuse discharge estimates from the profiles do not allow for spring flow and the likely very high leakage from the immediate spring surrounds. The overall diffuse discharge of the project area of approximately 20 by 70 km (1.4×10^9 m^2) at rates of 0.8 and 2.05 mm yr^{-1} are 1.1 and 2.9×10^6 m^3 yr^{-1}. Total spring flow is about 1000 m^3 day^{-1} or 3.65×10^5 m^3 yr^{-1} maximum (Kinhill Stearns, 1984; A.G.C., 1984). Holmes *et al.* (1981) showed that evaporation from the surrounds of the Emerald Spring, just west of Lake Eyre South, was considerably greater than the measured discharge of the well defined spring vent. They measured the extent of the shallow water table around that spring, and opined that the additional evaporation was supplied by "concealed leakage of water that cannot be gauged as a water stream". If leakage from

TABLE 6.3.1

Physiographic Units and Assigned Deep Aquifer Leakage Rates

Unit	Description	Area (relative)	Assigned leakage low mm yr^{-1}	high mm yr^{-1}
1	Low-lying areas associated with springs, creek-bed flood plains, and minor salt lakes, underlain by artesian aquifer	5 %	3.0	7.0
2	General low-lying gibber plain, underlain by artesian aquifer	60 %	1.0	2.5
3	Higher gibber plain, underlain by artesian aquifer	5 %	0.5	1.0
4	Areas where the aquifer is thin or absent by geophysical survey (Cockshell, 1988)	30 %	0.0	0.5

No allowance for springs:

	Overall average		0.8	2.05
	Sub-basin only average		1.1	2.7

Allowing for springs:

	Overall average		1.4	2.6
	Sub-basin only average		1.9	3.7
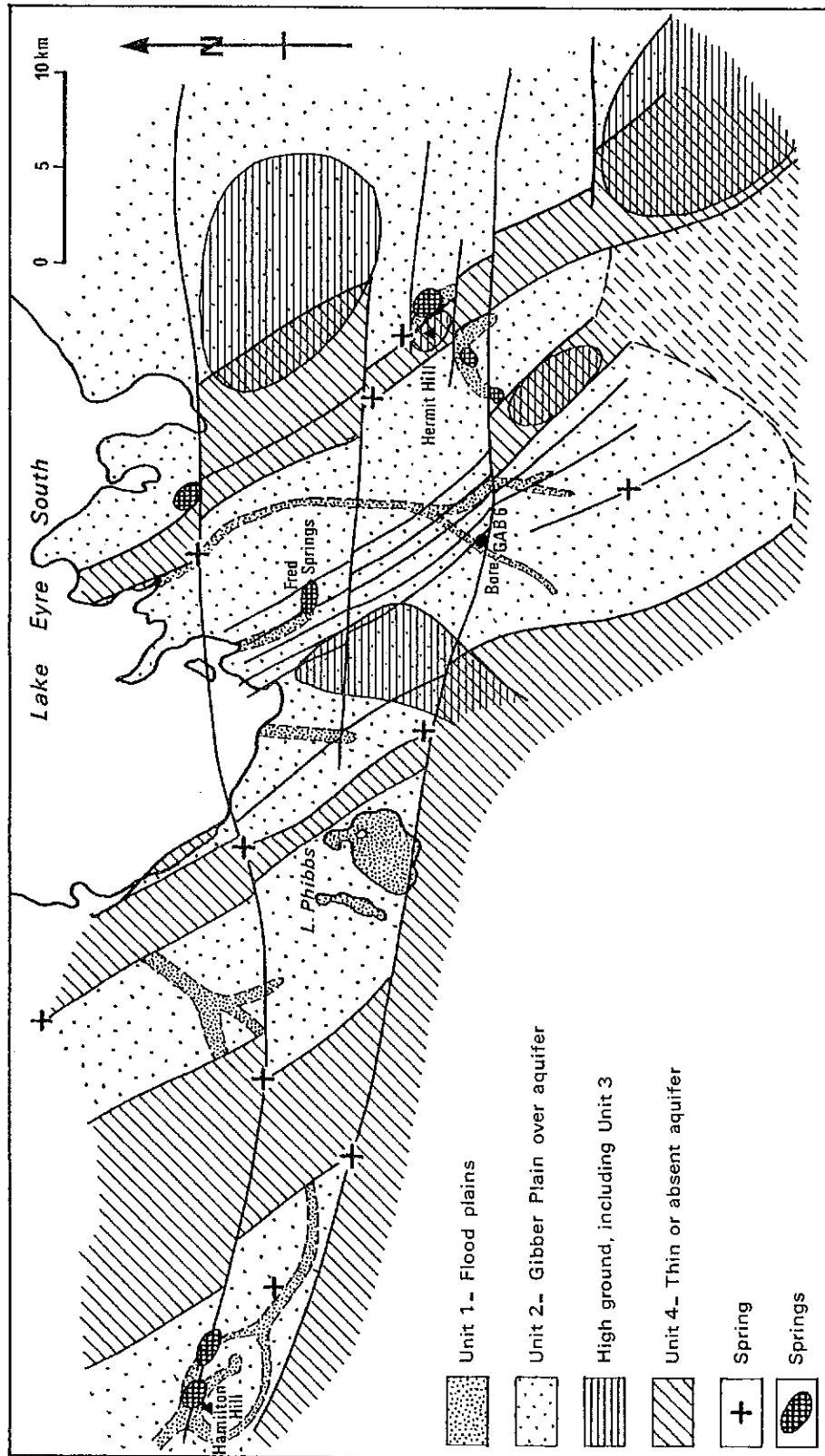
FIGURE 6.3.1 Physiographic Units for the Areal Estimation of Leakage in the Project Area.

Lake Eyre South

Hamilton Hill

L. Phibbs

Fred Springs

Bore GAB6

Hermit Hill

10 km

N

Unit 1 - Flood plains

Unit 2 - Gibber Plain over aquifer

High ground, including Unit 3

Unit 4 - Thin or absent aquifer

Spring

Springs

spring surrounds roughly amounts to the same as from the springs themselves, the total discharge (leakage) is between 1.9 and 3.6×10^6 m^3 yr^{-1}, implying overall averages of 1.4 and 2.6 mm yr^{-1} from the *total* area, or 1.9 and 3.7 mm yr^{-1} from the sub-basins. These are the figures that will be used in further discussion, and would not be greatly different even if the amounts of water estimated to be evaporated by springs and their surrounds were in error by up to 50% or more.

6.4 POSSIBLE IMPLICATIONS FOR THE ROXBY DOWNS WATER SUPPLY BOREFIELD A

The Olympic Dam Project's borefield A extends over approximately 10 x 30 km of sub-basin (nominated the south-west or borefield sub-basin). The original pre-exploitation leakage from this area is estimated to be between 5.7×10^5 m^3 yr^{-1} and 1.1×10^6 m^3 yr^{-1}, or 1560 and 3040 m^3 day^{-1}. This is considerably less than the amount of water that may be ultimately required from the borefield, 15 000 m^3 day^{-1} (Armstrong & Rowan, 1987). The influence of drawdown of the borefield in mid-1989 extended in a north-east direction under Lake Eyre South, and into the adjacent sub-basin, covering an area of about 30 x 20 km (Fig. 6.4.1, SADME, 1989), so that leakage is also "harvested" from those areas, perhaps doubling the contribution of this water source to say 3000 to 6000 m^3 day^{-1}.

It should be stated that the sub-basin and surrounds may well be able to sustain pumping rates higher than these leakage rates, as the increased hydraulic gradients in the aquifer itself will draw in water from further out in the basin, and there are possibly some other sources as well (Waterhouse & Armstrong, 1990). Because the original inflow into the main exploited sub-basin, as estimated from leakage, is quite below
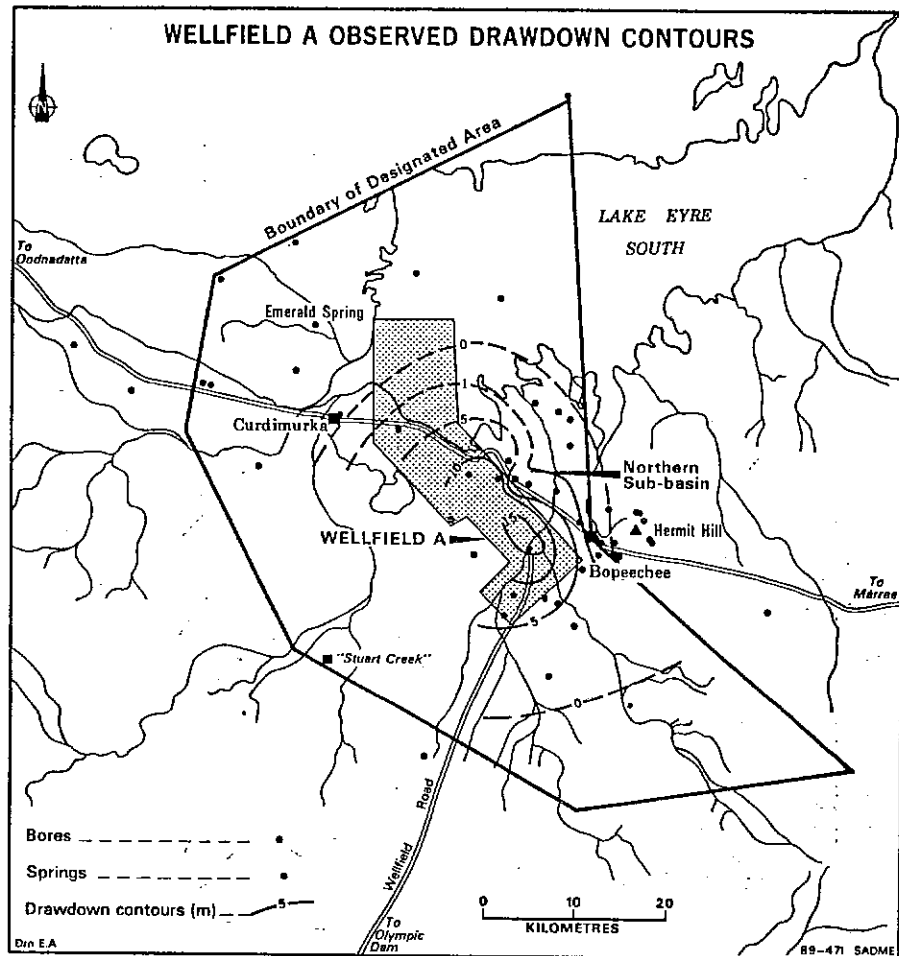
WELLFIELD A OBSERVED DRAWDOWN CONTOURS

FIGURE 6.4.1   Roxby Downs Water Supply Borefield A, Observed
Drawdown Contours mid 1989.   S.A.D.M.E., 1989.

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the demand placed on the borefield, it is very important to monitor
the piezometric relationship between the deep and shallow aquifers,
to guard against reversed leakage leading to increased salinity of
the water supply.   Such a monitoring program is in place, and the
borefield appears to be performing well to the end of 1989 (ibid.).

6.5 CONTRIBUTION OF GREAT ARTESIAN BASIN SALT TO LAKE EYRE

The origin of salts in Australian salt lakes has been a subject of
considerable debate over many decades. The four main sources are:
(a) oceanic salt from evaporated spray via rainfall and dust
fallout, known as cyclic salt,
(b) relic salt in marine or salt lake sediments, called connate
salt,
(c) dissolution of ancient evaporite deposits, and
(d) weathering products of catchment rocks.
Probably all four contribute in some cases, and all except evaporite
dissolution in most cases; the argument is as to which is the main
source. It is beyond the scope of the scope of this work to go into
matter in depth, but a discussion in the context of Lake Frome is
given by Draper and Jensen (1976).

This section presents some preliminary "back of envelope"
calculations regarding the contribution of GAB salt to Lake Eyre.
Not enough data were obtained to assign even an estimate of
confidence to the presented figures, and in this context no attempt
is made to verify the assumptions made.

Leakage of Great Artesian Basin waters must provide a certain
quantity of salt to Lake Eyre, particularly the southern lake.
Bonython (1956) calculated that spring flow would provide all of the
halite crust then found in Lake Eyre North in about 25 000 years,

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but points out that other sources are more important.

The total pre-exploitation loss of GAB water in the project area is estimated above to be between 1.9 and 3.6 x 10^6 m^3 yr^{-1}. Using 2.5 x 10^6 m^3 yr^{-1} for the sake of calculations, and taking chloride content to be 1 g l^{-1}, as did Bonython, some 2.5 x 10^6 kg yr^{-1} of chloride, or 4.1 x 10^6 kg yr^{-1} of Na Cl is added to storage in the Bulldog Shale and overlying soil. As a long-term average, the same amount may be said to be washed into Lake Eyre South, by lateral shallow groundwater flow, base flow in creeks, or in surface water that has dissolved efflorescent salt on top of the soil surface or eroded top soil and the salt within it. Note that this assumes windblown losses are balanced by gains.

The project area makes up about a third of the total part of the catchment of Lake Eyre South that is underlain by artesian pressure parts of the GAB, so the total GAB contribution to the catchment is say 12.3 x 10^6 kg yr^{-1}. The salt crust in Lake Eyre South contained 3 x 10^{10} kg of halite prior to the 1989 flood (Allen et al., 1986), and this is largely the remnant of salt that flowed into the lake from Lake Eyre North in earlier floods. Using that amount simply as a comparison, it would be equivalent to 2 400 years of GAB contributions. The contribution of GAB derived salt to Lake Eyre South, at least the underlying brines if not the crust, does have scope to be significant, as it has probably occurred for many millennia. This potential importance as a source is because of Lake Eyre South's relatively small catchment area, much of which is underlain by part of the discharge area of the GAB.

The contribution of GAB salt to the full Lake Eyre complex is less important. Lake Eyre North has an enormous catchment area over which

to catch cyclic salt and to provide salt by the weathering of rock. While there may be a contribution of salt to the GAB aquifer from older, underlying basins such as the Adavale Basin, which is known to contain evaporite deposits (Wells, 1980), the water is of meteoric origin as evidenced by its stable isotope composition. Connate salt has probably been long flushed from the aquifer itself, and weathering of its minerals may contribute some sodium but little chloride. Cyclic and some weathered salt is taken in with recharge waters, but the recharge area is much smaller than the surface water catchment of Lake Eyre, so the part of the total salt budget travelling by the underground route to the Lake is small.

The speculation so far regards the equilibrium situation. The amount of salt stored in the shallow groundwater brines and soil profiles around the lake is very large, and increased shallow aquifer hydraulic gradients, or rapid erosion of soils could lead to a much greater than average contribution of GAB originated salts into the lake on the time scale of hundreds to a few thousand years.

CHAPTER 7

SUMMARY AND CONCLUSIONS

7.1 SUMMARY

7.1.1 GENERAL

This project was inspired by the opening of the margin of the Great Artesian Basin to large-scale commercial water extraction, and the need to have a measure of an aspect of the water balance of the region that had never been measured directly in the past, yet was a major factor concerning the successful long-term use of the resource. Directly measuring diffuse leakage from the Great Artesian Basin gives substance to the process that accounts for much more of the natural discharge of the resource than the famous mound springs, yet had never been measured at all, but only estimated by rather indirect means. The usefulness of such measurements to contribute to establishing a long-term, sustainable water supply for a major resource development gave the project a more immediate application and relevance as well.

7.1.2 THEORETICAL DEVELOPMENT

The convection – diffusion method used to estimate the evaporatively driven upward flux of water is a fairly recent innovation. Theoretical development and laboratory verification of this approach to estimating evaporation using stable isotope profiles (Zimmermann *et al.* 1967; Barnes & Allison, 1983) was followed by a successful field application in an

Australian dry salt lake bed (Allison & Barnes, 1985). Some theoretical advances followed, covering non-isothermal and non-steady-state conditions (Barnes & Allison, 1984; Walker *et al.*, 1988). Scant attention was paid to better understanding one of the least well known and uncertain parameters involved in calculations, the impedance factor, that allows for the effects of geometry, interaction with the solid particles of the soil, and the contribution of vapour flux to the overall water movement.

This work draws attention to the most complete and accurate measurements of impedance factor for the diffusion of water isotopes in soils, those by Nakayama and Jackson (1963) and Scott and Paetzold (1978). They show that the impedance factor for total water flux increases at low water contents, due to the increasing contribution of vapour movement. The more detailed work of Nakayama and Jackson (1963) shows that a maximum value is reached at a certain water content, and that below this the impedance factor drops rapidly; this is due to adsorption of water onto soil particle surfaces. The shapes of the impedance factor versus water content curves determined by these two works is adequately explained by existing theory, providing the contribution of water vapour is properly included. The relationship for an individual soil types varies, as might be expected due to mineralogical and particle size distribution differences. It is not possible to use the small number of precise measurements to give a predicted relationship for other soils and materials. Rather, the impedance factor is perhaps best considered to be roughly constant at 0.5 between volumetric water contents between about 0.1 and 0.5 m^3m^{-3}, with an uncertainty of about 40%.

Many measurements are available in the literature of the impedance factor for the diffusion of chloride in soils, which is likewise

dependent on water content and soil type. Using data obtained experimentally, and compiled from the literature by Johnston (1987), the relationship of $f_l = 0.8 \ \theta_v - 0.05$ was chosen for use with the field profiles of this project, which are mostly fine-grained. This choice is again somewhat subjective, and may have a similar uncertainty to that chosen for isotopes of water.

7.1.3 LABORATORY TECHNIQUE DEVELOPMENT

The investigation into extraction techniques required before the field study could proceed demonstrated both the wide applicability as well as some of the limitations of the approach using azeotropic distillation with hydrocarbons. The "gypsum problem" motivated the investigation of a number of possible low-temperature alternatives to extract pore water from porous media for isotopic analysis. Those investigated were azeotropic distillation with hexane, a mass balance approach, vacuum distillation at room temperature, and the adsorption of vapour by desiccants. The accuracy of azeotropic distillation with toluene and kerosene from non-gypseous materials was also investigated in some depth.

It is concluded that azeotropic distillation is a suitable method to extract water from porous media for isotopic measurement. Toluene gave the best results for distillation of waters, and kerosene for non-gypseous soils, as well as the hydrated mineral gypsum by itself. The errors are about 1.5 and 0.2 permil for δD and $\delta^{18}O$, respectively, although greater accuracy is obtainable with "easy" materials, notably sand at moderate to high water contents. Distillation with hexane, restricted to two to three hours duration, gives the most representative value for pore water from a gypseous medium. The isotope values obtained

by such a hexane distillation are systematically light by 3 and 1.1‰ for δD and $\delta^{18}O$, with errors of about 2 and 0.3‰, respectively. The other techniques investigated proved unsuitable.

An international, interlaboratory comparison of various methods to determine the stable isotope composition of three soils, one with two moisture contents, showed an alarming spread of results. The overall scatter of the data is much greater than that claimed for any of the methods individually, and appears to be both material and water-content dependent. The intercomparison highlights the need for careful laboratory technique, and verification of methods for different soil types. Comparisons of the absolute values of soil water isotope data obtained by different laboratories should be made with caution when different techniques are used.

7.1.4 FIELD RESULTS AND APPLICATION

The relationship between deep and shallow aquifer waters in the project area was fairly clearly expressed by their isotopic composition and major ion chemistry. The fresh to brackish deep waters from the main GAB aquifer are generally of the $Na-HCO_3-Cl$ hydrochemical type, although Cl becomes the dominant anion in the western part of the project area. Salinity varies from about 1900 to 5100 mg l^{-1} total soluble solids (TSS), being higher in the west and near the very edge of the basin. The average stable isotope composition is about -50‰ δD and -7.5‰ $\delta^{18}O$, lighter than the GAB average but still falling on the world meteoric line.

Shallow ground waters are saline to hypersaline, with salinities varying between about 37 000 and 200 000 mg l^{-1}. Their chemistry is strongly dominated by Na Cl, with secondary sulphate in most cases,

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occasionally accompanied by high HCO_3. The stable isotope compositions of shallow waters fall on a line on a $\delta^{18}O$ – δD plot described by the equation

$$\delta D = 3.8\ \delta^{18}O - 22.6$$

that passes through the value of the deep water. This relationship is consistent with water coming from the deep source and being isotopically enriched by evaporation through a dry top soil. The chemistry of the water is changed during passage through the shale aquitard, by evaporative concentration and deposition of evaporite minerals, and recycling of the more soluble salts in the unsaturated zone.

An extreme rainfall event occurred during the project, permitting some observations regarding recharge to the shallow aquifer and contribution of shallow groundwater to the base flow of local, ephemeral streams. Deep infiltration leading to significant recharge did occur, but appears to have been very patchy in its effectiveness. Low lying areas, where rain water ponded, appear to have been sites of high and immediate recharge, whereas higher areas of the landscape that shed most of the rain showed evidence of very limited infiltration and negligible or zero recharge. The shallow aquifer received enough water from the recharge event to raise the water table far enough to provide a saline base flow to the deeper incised creeks in the area. These creeks were still flowing a month after the last significant rainfall, with electrical conductivities between about 25 000 and 50 000 $\mu S\ cm^{-1}$, equivalent to TSS between about 15 000 and 30 000 mg l^{-1}. Heavy isotopes were enriched in the creek flows showing the influence of the shallow groundwater. The water in Lake Eyre South was less saline than any of its then-flowing tributaries, and had an isotopic composition lying very close to the meteoric line, suggesting the vast majority of the water came from the main flooding event due to heavy rain.

The holes drilled could be divided into two main groups by their location and lithography, those on flood plains and those on gibber (stony) plains. The flood plain holes at Hermit Hill in the east, and all of the gibber plain holes passed into the weathered top part of the underlying sedimentary rock the Bulldog Shale, that is the aquitard overlying the GAB aquifer in the Algebuckina Sandstone / Cadna-owie Formation some 30 to 150 m below.

Following successful laboratory technique development, it was possible to extract gypsum crystal water and pore water from the same soil sample. Gypsum water was not in isotopic equilibrium with the coexisting pore water. Rather, its composition lay on the same evaporation line on a $\delta^{18}O$ - δD plot as the pore waters, but at the lighter end of the span of pore water values. This is interpreted as probably indicating that the bulk of the gypsum was deposited in the past, either when the ground surface was at a higher elevation before being removed by erosion, or deposited when the climate was less arid and the evaporative enrichment of pore waters was not as great as at present.

The chloride profiles of 11, and deuterium profiles of 10 holes could be interpreted by steady-state convection-diffusion theory to yield estimates of evaporation. Profiles were chosen that showed good evidence to be consistent with the assumptions of the theory, especially steady state. Evidence after the heavy March 1989 rain event suggests that such slightly elevated sites were undisturbed by infiltration except for the top half a metre or so. The adopted evaporation estimates, with an approximate uncertainty of 50%, ranged from 0.5 to 4.5 mm yr^{-1}. Where $\delta^{18}O$ was examined as well as δD in the profiles, the data fitted evaporation lines with slopes of between 2 and 3, and in most cases intercepting the meteoric line above (heavier values) the average deep groundwater value.

The evaporation estimates are consistent with the less certain, independent calculations of water vapour flux at the top of some holes, and the semi-independent estimate from the depth of the isotope maximum. The estimates are in rough accord with most of those obtained in arid sites in North Africa by other workers using similar methods. They are somewhat less than the leakage estimated hydrodynamically, which suggested rates of between 14 and >68 mm yr^{-1}. The hydraulic calculations depend on a very poorly known parameter, the vertical hydraulic conductivity of the Bulldog Shale, and so may be considered very rough calculations.

The point estimates of evaporation may be taken as close indicators of leakage from the deep aquifer to the water table. An areal estimate was made by dividing the region into four physiographic associations based on topography and the distribution of the aquifer. After allowing for the contribution of natural springs and their surrounds, the areal average pre-exploitation leakage (discharge) is estimated to be between 1.4 and 3.7 mm yr^{-1} over the entire project area, or between 1.9 and 3.7 mm yr^{-1} taken over areas underlain by a significant artesian aquifer, or "sub-basins".

The area of drawdown of the Olympic Dam Project borefield A is approximately 600 km^2, so that it is "harvesting" a previous leakage of between about 3000 and 6000 m day^{-1}. The area appears to be able to support a long-term pumping rate of at least 9000 m day^{-1}, most of the additional water being drawn into the borefield by the increased hydraulic gradients induced by the pumping. The drawdown in the borefield is generally not excessive, but small local reversals of hydraulic gradient between the GAB aquifer and water table in the immediate vicinity of production bores are being monitored, as extensive reversal could lead to eventual salinisation of the aquifer.

7.2 SCOPE FOR FURTHER WORK

Any research programme will usually produce as many questions as it answers, if not more. Some aspects of this work were per force only briefly touched upon, and even the areas investigated in depth leave scope for further investigation and refinement.

There are several processes that require further investigation with regard to their effects on convection and diffusion of solutes in porous media, as well as the movement of water generally. This constitutes much of the field of soil physics, a large subject and the scene of much research world-wide. Of particular interest to studies such as this one are the effects of anion exclusion, non-participating water, matric suction and osmotic effects on diffusion. The understanding of non-steady-state processes, and the approach to steady-state, require continued research and application.

The convection-diffusion approach to estimating evaporation has had a very limited application to date, because of one very major drawback – its restriction to non-vegetated, arid-zone sites. Its usefulness in more temperate, vegetated areas is yet to be demonstrated. Interaction with other techniques and disciplines may lead to a broader application, as one of a range of methods that each tell part of the story and need to be applied together, rather than the virtual "stand alone" method used to date. Indeed, comparison with other methods, the conventional estimation of leakage by hydraulic calculation or local groundwater flow modelling by computer (A.G.C., 1984), and the less common calculation of vapour flux from matric and osmotic suction data, has been very useful in this study, as was the comparison of isotopic and chemical data between the deep and shallow groundwaters.

Experimental data of this work and the interlaboratory comparison showed the need for careful investigation and practice of such a basic and routine procedure as procuring a water sample for isotopic analysis from naturally occurring soils and rock material. The dissemination of such results by laboratories, by publication as technical reports available to other workers in the field, would be of mutual benefit, and probably lead to more uniform and comparable results between groups. The use of desiccants to adsorb equilibrated vapour from a sample of porous material would seem a possible candidate for further work, and is potentially useful with all porous media.

More point estimates of evaporative discharge of groundwater have been obtained by this project than any previously published studies that could be found. The perennial problem of scaling up such a spatially variable quantity occurs as it does with any similar work. One possible manner of extending the estimates might be shallow geophysical techniques, such as are being developed to map soil salinity and even recharge (Williams & Baker, 1982; Cook et al., 1989). Another method could be based on air- or satellite- borne remotely sensed data, as was attempted by Menenti (1984) and Schmid (1985). Both approaches require calibration or "ground truthing", perhaps best supplied by solute profile modelling, as here, or by micrometeorological methods.

In the field area itself, the high leakage of pressurised water in the immediate vicinity of mound springs was not measured but could only be guessed at. The springs mostly exist is areas where the confining bed to the aquifer is fractured, and vary from strongly flowing streams issuing from the ground, through pools and vegetated mounds to merely damp patches of soil. The next gradation in the series is the "almost" spring, where leakage is much greater than the one to a few millimetres per year of most of the margin of the basin, but not strong enough to

show at the surface. A watertable sustained only half a metre or less beneath the ground surface could represent a leakage and evaporation of tens to hundreds of millimetres per year, perhaps over large areas between individual springs.

The extent and magnitude of lateral flow in the more permeable parts of the upper Bulldog Shale is another matter that may be worthy of further investigation. The contributions of leaking artesian water and the rare surface recharge events would be difficult to distinguish, but at least the contribution of shallow groundwater and salt into Lake Eyre could be estimated. Of more economic importance is the potential contamination of useful GAB water with salt from the shallow aquifer, if in the future the drawdown caused by the Roxby Downs borefield A causes extensive reversal of flow between the two groundwater bodies.

On a more general note with regard to the Great Artesian Basin, the possibility that gypsum water has recorded evaporative isotope profiles in areas such as near Coober Pedy deserves further investigation. The two holes of this project sited in likely terrain were not drilled to sufficient depth for the concept to be checked. The isotopic differences in gypsum water and soil water, which might indicate greater burial depths (higher ground elevations) or less arid climates in the past, merit further investigation. One curious phenomenon noted, but not investigated at all, is the peculiar variation of natural spring flow that occurs in the mound springs of the region. The cause(s) apparently still remain a puzzle for the Olympic Dam Project officers who monitor the flows, and would make an interesting study, if a somewhat esoteric one.

7.3 CONCLUSIONS

The convection-diffusion method of estimating groundwater evaporation proved a useful tool to obtain the first direct measure of the major natural discharge process of Australia's largest groundwater basin. The estimates provide an independent measure of the safe yield of a section of the margin of the Great Artesian Basin that has been recently developed as the main water supply to a major mining venture and its associated township. The existing borefield appears to be operating successfully, but may be close to the long-term sustainable yield of the part of the basin that is being utilised. The position of the borefield within the basin is at perhaps the most appropriate location possible, as a large proportion of its supply comes from "harvesting" water that was previously lost to the atmosphere as evaporation.

The development of isotopic and chloride techniques for discharge and recharge studies is at a point where they must be combined with more conventional soil physical and other approaches to enable wider application to hydrological problems in a wider range of physical environments. It is also appropriate to consolidate the basis of these techniques by more careful verification of the apparently simple laboratory procedures that are involved, and to measure the important, yet poorly known parameter needed for practical calculations, the impedance factor for diffusion of chloride, and particularly that for isotopes of water.

Overall the project must be considered a success, as:

* the theory was extended to include a shallow boundary condition,

* the value and error of the parameter most limiting to the accuracy of the method has been better defined than previously,

* the obstinate problem of extracting pore water from gypseous samples was surmounted,

* the problem of poor intercomparability between different laboratories' extraction techniques has been brought to light,

* a useful estimate of pre-exploitation leakage from a commercial borefield that uses that leakage provided;

* these are the first direct measurements of a major component of the water balance of a very large and important aquifer system, and

* avenues for further research and application have been opened.