CHAPTER
9

THERMAL REGIMES IN WETLAND SEDIMENTS

9.0 INTRODUCTION

This chapter investigates and expands upon the difficulties inherent in directly measuring the sediment heat flux and discusses its importance in the thermal balance of shallow wetlands. Concepts of conducted and advected heat flow are introduced and applied to the Perry Lakes data. The thermal patterns in East Lake sediments are presented and used to demonstrate how these reflect daily and seasonal changes in lake-sediment and lake-aquifer interaction.

Wetlands, as the exposed portions of unconfined aquifers, may be conveniently thought of as 'windows on the water table'. Similarly they are also 'windows to the sun'. They intercept solar energy and there is a net transfer of some of this energy into the aquifer. In this sense shallow lakes and wetlands are essentially 'thermal sumps'. Within lake basin sediments, thermal patterns are the combined result of surface water-groundwater interaction and diurnal, seasonal, and even longer term changes in lake water temperature.

Below the land surface, where the vadose zone is relatively thin, solar energy is conducted into the aquifer through the soil. The effective perturbation depth of surface temperature fluctuations is only about 10m (Lovering & Goode, 1963 cited Domenico & Schwartz, 1990). Within the upper 10-20m groundwater temperature may be 1° to 2° higher than the mean local annual temperature. These surface effects are superimposed on the regional geothermal gradient. Below the range of these surface perturbations, there is a steady temperature increase where geothermal heat is conducted upwards from the earth into the aquifer. Our attention is on lakes and *Qse*, the net heat loss through the upper surface of the lake sediments.

Qse is made up of solar energy absorbed directly on the upper surface of the sediments and vertically advected and conducted heat fluxes across the water-sediment interface. The general flux/storage balance predicts that this quantity must be made up from depleted storage within the sediments and heat advected and conducted from the lower surface of the sediments, including geothermal heat from below.

Thermistor strings were installed into the East Lake sediments over winter 1996. The proposed approach was to use these point source data to extrapolate over the entire lake basin employing one or more of the theoretical approaches outlined in Section 9.1. Manipulation of the early data, however, augmented by manual measurements at other points in the lake plus sun and shade distribution observations, quickly pointed to the necessity for a more holistic approach. A carefully executed thermal balance in which all components were individually measured including evaporation (but excluding sediment flux) appeared to be the method most likely to integrate the complex sediment thermal regime. This however necessitated a direct and totally independent measurement of evaporation and lead directly to the design and installation of the floating evaporation pan in December 1996 (Chapters 5 & 8).

9.1 THERMAL REGIMES IN WATERTABLE LAKE SEDIMENTS

9.1.1 Concepts of conduction and advection

In wetlands heat is both conducted and advected. Conduction requires a temperature gradient (Pitts & Sissom 1998). Conduction is a linear process described by Fourier's law (Incropera & DeWitt 1996)

$$H = -\kappa \Delta T \tag{9.1}$$

where H is the heat flux (as heat per unit area), T is temperature, ΔT is the temperature gradient and κ is thermal conductivity, a proportionality constant linking the two. In the present work, the term 'conduction' includes molecular transfer and any 'turbulent' transfer[1] of sensible heat. Diffusion is a more accurate description of this process. Diffusion, as described by Fick's first law (Bird *et al* 1960) also takes the form of equation (9.1). Saturated sediments are essentially two phase systems comprising mineral grains and interstitial water. Within them conducted heat is influenced by the physical characteristics of both.

The thermal conductivity is the quantity of heat transmitted per unit time through a unit cross sectional area under a unit temperature gradient. It considers the volume fractions and conductivities of the solid and liquid phases. It is influenced by many factors including grain size and shape, nature of grain to grain contacts, pore size, porosity, grain specific gravity, grain thermal conductivity, degree of saturation and salinity of the pore water (Lapham 1989, Domenico & Schwartz 1990) as well as direction and, perhaps, the temperature gradient itself. In a saturated sediment, the effective thermal conductivity κ_e

[1] Such 'turbulent' exchange is often referred to as convection

is used. The divergence ∇ of the conductive heat flux is the rate of loss of heat per unit volume due to diffusive transfer

$$\nabla(-\kappa\nabla T) \tag{9.2}$$

Advection is direct flow of heat with the flow of water. When heat energy moves with water, the advective flux of the water is $\rho\upsilon$ and the advected heat is $\rho c\upsilon T$ where c is the specific heat, ρ is the density and υ is the Darcy velocity. The corresponding divergence and rate of loss of heat by advection is

$$\nabla(\rho c\upsilon T) \tag{9.3}$$

In the partial differential equation of heat balance, equations 9.2 and 9.3 are combined to include the two fluxes, conduction and advection (Bird *et al* 1960). This describes constant flow in the vertical direction for constant κ, ρ and c.

$$\frac{\kappa}{\rho c}\frac{\partial^2 T}{\partial z^2} - \upsilon\frac{\partial T}{\partial z} = \frac{\partial T}{\partial t} \tag{9.4}$$

The right hand term derives from $\rho c\ \partial T / \partial t$, the net rate of accumulation of heat in the sediment volume. Equation 9.4 describes the continuous heat balance in vertical flow with conduction, groundwater discharge, lake water recharge and storage in the sediments.

Heat capacity varies little among mineral solids that typically make up fine and coarse grained sediments. It does however vary considerably depending on bulk density. Sediment thermal behaviour is further influenced by the thermal conductivity and heat capacity under unsteady thermal conditions. Thermal diffusivity $\kappa / \rho c$ is the ratio of thermal conductivity to volumetric heat capacity. A sediment of high thermal diffusivity will change temperature rapidly, in response to a sudden external temperature change. The thermal diffusivity of saturated fine and coarse grained sediments also varies with sediment bulk density.

In wetlands the linked advective (fluid) and conductive (diffusive) components may oppose or augment each other. Figure 9.1 illustrates three simple cases (left to right) where fluid and conductive components oppose each other, augment each other or combine in complex relationships where some or all of the fluxes can have both horizontal and vertical components. Not shown, but implicit in each diagram is a heat storage component in the sediments.

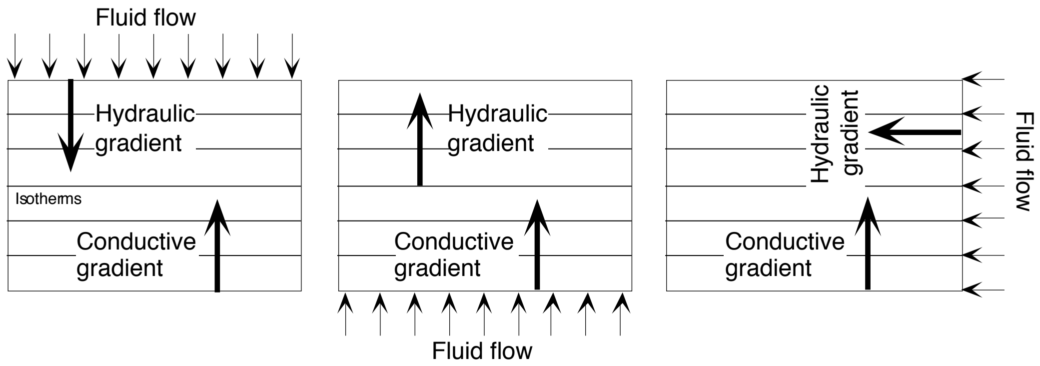


Figure 9.1 Idealised cases of heat advected in fluid flow superimposed on conductive gradients (Figure adapted from Domenico & Schwartz 1990). In real world situations all the fluxes are likely to have both vertical and horizontal components.

9.1.2 Concepts of daily and seasonal variation

Figure 9.2a shows the principal components of the East Lake thermal balance. The solar terms *Qs* and *Qa* are particularly important. The clear, shallow water and extremely high surface area to volume ratio (Figure 9.2b) which varied in 1996-97 from about 4:1 to 40:1 ensures the importance of the solar energy fluxes. The sediments are heated through contact with the water column and directly by solar insolation.

Diurnally, heat is conducted in and out of the sediments. Surface, mid level and bottom water temperatures were monitored in the centre of the South Basin. Close to the dividing stream line lake-aquifer water flows are minimised and conduction between the water column and the sediments and solar insolation predominate. The direction of conducted heat transport varies during the day. In Figure 9.3 the daily temperature difference between bottom water and mid level water is plotted for 06:00 and 15:00 hours. On most days, at 06:00 bottom water is warmer than mid level water. The sediments are also warmer than the water column and heat is conducted from them into the water column which has cooled radiatively over night. During the day solar energy heats the water column so that by mid afternoon the mid level water is almost always warmer than the water-sediment interface. Heat is now conducted downwards into the sediments.

We can now consider the entire lake basin and the combined effects of advected, conducted and radiated heat. Advected heat flow varies spatially and temporally depending on the prevailing flow regime. Schematic cross sections in Figure 9.2c summarise daytime and night time fluxes during flow-through and recharge flow regimes. Conductive and advective fluxes may augment or oppose each other depending on the time of day, location within the lake basin and the prevailing flow regime. The net daily sediment flux also varies from negative (into the sediments) to positive (into the water column). At Perry Lakes the lake lining sediments are up to 3m thick in comparison to the average water depth of 0.2 to 0.5m. They have both a much greater volume and heat capacity than the water and therefore a significant potential to store heat.

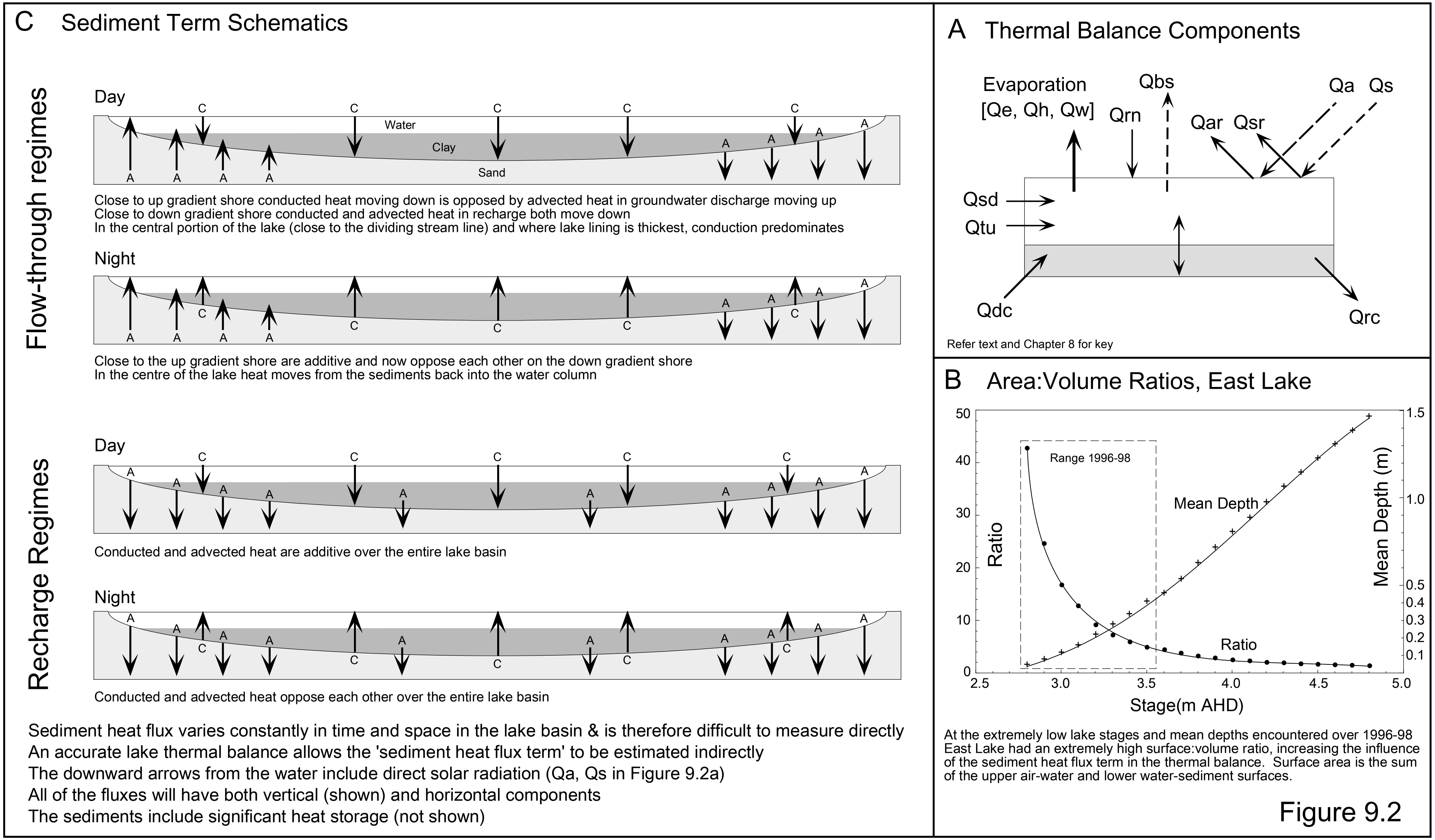

C Sediment Term Schematics
Flow-through regimes
Day
Water
Clay
Sand
Close to up gradient shore conducted heat moving down is opposed by advected heat in groundwater discharge moving up
Close to down gradient shore conducted and advected heat in recharge both move down
In the central portion of the lake (close to the dividing stream line) and where lake lining is thickest, conduction predominates
Night
Close to the up gradient shore are additive and now oppose each other on the down gradient shore
In the centre of the lake heat moves from the sediments back into the water column
Recharge Regimes
Day
Conducted and advected heat are additive over the entire lake basin
Night
Conducted and advected heat oppose each other over the entire lake basin
Sediment heat flux varies constantly in time and space in the lake basin & is therefore difficult to measure directly
An accurate lake thermal balance allows the 'sediment heat flux term' to be estimated indirectly
The downward arrows from the water include direct solar radiation (Qa, Qs in Figure 9.2a)
All of the fluxes will have both vertical (shown) and horizontal components
The sediments include significant heat storage (not shown)
A Thermal Balance Components
Evaporation
[Qe, Qh, Qw]
Qrn
Qbs
Qar
Qsr
Qa
Qs
Qsd
Qtu
Qdc
Qrc
Refer text and Chapter 8 for key
B Area:Volume Ratios, East Lake
Range 1996-98
Mean Depth
Ratio
Mean Depth (m)
Stage(m AHD)
At the extremely low lake stages and mean depths encountered over 1996-98 East Lake had an extremely high surface:volume ratio, increasing the influence of the sediment heat flux term in the thermal balance. Surface area is the sum of the upper air-water and lower water-sediment surfaces.

Figure 9.2

9.2 SEDIMENT HEAT FLUX TERM

9.2.1 Different approaches to measurement

Given the complex nature of heat movement in lake-aquifer systems, many different approaches have been used to quantify the sediment heat flux *Qse*.

Simple conductive gradients

The simplest expressions for *Qse* occur in many deep lakes and high latitude lakes where there is effectively little or no diurnal or seasonal change in bottom water temperature. Here Fourier's law (refer equation 9.1) can be applied to describe conduction across a temperature gradient

$$Qse = \kappa\left(T_2 - T_1\right) \tag{9.5}$$

where

κ sediment thermal conductivity

T_1 temperature at the water-sediment interface

T_2 temperature at some depth within the sediments

Examples include arctic lakes (Gibson *et al* 1996), high latitude wetland complexes (Mendez *et al* 1998), and temperate lakes (Likens & Johnson 1969). Similarly Hondzo *et al* (1991) estimated sediment heat flux in a Minnesota lake of similar size to Perry Lakes using temperature time series at the sediment water interface and within the sediments down to 1.5m below the interface. Heat flux was calculated as the rate of change in sediment heat storage obtained by integrating sediment temperature profiles $T(z,t)$

$$\text{sediment heat flux} = \rho_s c_{ps} \frac{\partial}{\partial t}\int_0^{zd} T_s(z,t)\,dz \tag{9.6}$$

where

ρ_s bulk sediment density

c_{ps} sediment specific heat

Diurnal and seasonal fluctuations

Thermal conditions in shallow lakes and rivers which freeze over winter are highly dependent on heat exchange between the water and sediments (Pivovarov 1973). Theoretical calculations predict an almost equal balance between heat lost into the sediments over summer and heat returned to the water column from the sediments over winter (Pivovarov 1973, Fig 9). In such lakes, heat stored in bottom sediments is an important source of winter heat.

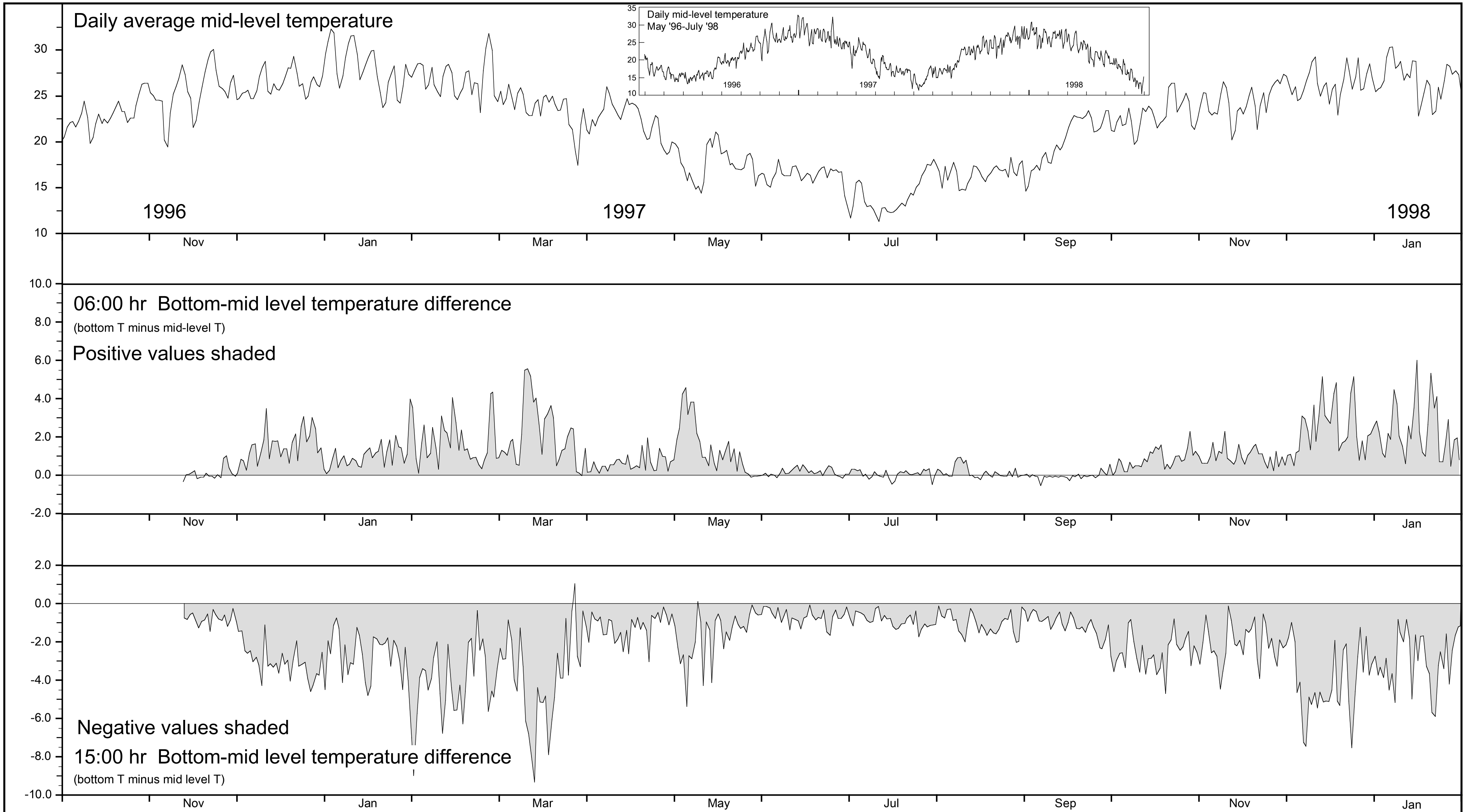

Daily average mid-level temperature
Daily mid-level temperature
May '96-July '98
1996
1997
1998
30
25
20
15
10
Nov
Jan
Mar
May
Jul
Sep
06:00 hr Bottom-mid level temperature difference
(bottom T minus mid-level T)
Positive values shaded
10.0
8.0
6.0
4.0
2.0
0.0
-2.0
Negative values shaded
15:00 hr Bottom-mid level temperature difference
(bottom T minus mid level T)
-4.0
-6.0
-8.0
-10.0
All y axes: Temperature °C

Figure 9.3

Sediments below temperate lakes typically display a sinusoidal temperature cycle, characterised by a decrease in amplitude and phase lag with increasing depth below the water-sediment interface. Simple harmonic functions can be used to describe this oscillation (Likens & Johnson, 1969). At Williams Lake, Sturrock *et al* (1992) were able to measure the sediment term directly using an equation described by Pearce and Gold (1959) where the heat flux at time t and depth x is defined by

$$Qse = akT_s e^{-ax} \sin\left(\omega t + \phi + \frac{\pi}{4} - xa\right) \tag{9.7}$$

where

$a = \sqrt{\dfrac{\pi C_v}{Pk}}$

C_v volumetric heat capacity

k thermal conductivity

P period of temperature variation

$\omega = \dfrac{2\pi}{P}$

T_s amplitude of temperature variation at the surface

ϕ phase lag in time of temperature variation at the surface

Hughes (1967) used a similar technique to measure heat flux through the bottom of the Salton Sea, California. Tsay *et al* (1992) used a harmonic analysis of temperature oscillations at the water-sediment interface to model sediment heat flux in small lakes in New York state. Walker (1973) used a similar approach to estimate sediment heat storage for Lake Werowrap (Victoria) using an expression derived by Neumann (1953, cited Walker 1973). The expression describes the sediment term at time t

$$Qse = \frac{c_m}{\sqrt{2b}} a \sin(A + \omega t - \pi / 4) \tag{9.8}$$

where

a and A are respectively amplitude and phase angle of the sin wave

c_m is the thermal capacity

$\omega = 2\pi / T$, T being the wave period (1 year)

$b = \sqrt{\omega / 2K}$

K is the sediment diffusivity

The expression defines the case where the bottom water temperature differs from the sediment temperature by a negative phase angle of π/4 radians (*i.e.* the sediment temperature lags behind the bottom water temperature by three months).

Temperature oscillations can also be used to estimate other parameters. Stallman (1965) demonstrated that in homogeneous sediments, rates of vertical non isothermal recharge from surface water bodies can be estimated using the attenuation of sinusoidal fluctuations of temperature with depth. This concept was extended by Bredehoeft & Papadopulos (1965) who used vertical groundwater temperature profiles to determine vertical groundwater velocity and hydraulic conductivity. Lapham (1989) applied this to streams in the eastern United States, to estimate recharge rates, discharge rates and vertical hydraulic conductivity. Similarly Hunt *et al* (1996) used sediment temperature profile modelling to estimate groundwater discharge and recharge in a wetland complex in Wisconsin.

9.2.2 Approach adopted at Perry Lakes

In Chapter 8 we noted that shallow lakes are not generally considered to be good candidates for thermal balance studies. In such lakes large proportions of the incoming day time radiation are conducted to both the water and underlying sediments. At night much of this energy may be lost as long wave radiation. Shallow lakes are incredibly dynamic systems when considered from a thermal perspective.

At Perry Lakes these problems were compounded by large area/volume changes which occur on both daily and seasonal time scales. East Lake volume and area may double or triple within a few hours in response to storm water and top up. Throughout the day shade from fringing trees can affect large areas of the lake. This is particularly so in winter with low sun angles. In short, such lakes present gross spatial and temporal complexities.

Ten thermistor strings were installed over winter 1996 (Figure 5.1a). As noted previously, we anticipated being able to use these ten point data sets to extrapolate over the entire lake on a daily basis using one or more of the approaches outlined in Section 9.2.1. Trial manipulation of the initial thermistor data augmented by manual measurements at other points in the lake indicated that a more holistic approach would be necessary. In particular, daily observations of large changes in lake area over small changes in lake stage (Figure 9.2b) and extensive shading (particularly at low winter sun angles) were identified as problems which would be difficult to resolve using a few fixed data collection points and led directly to the thermal balance approach. Any balance in which one component is derived as a residual is a compromise because the cumulative errors in measuring all the other components are reflected in the residual (Winter 1981). In this case the shear magnitude of the sediment flux term suggested that the seasonal patterns of sediment heat flux would still be defined even if their quantification contained some error.

9.3 SEDIMENT HEAT FLUX IN LAKES AND WETLANDS

9.3.1 Literature review

Thermal balances on water table dominated rivers in Britain (Table 9.1) showed that conduction into the river bed was the dominant form of non advective heat loss, while groundwater discharge was a significant contributor of advected heat gain to the system. Shallow rivers are similar in many respects to very shallow lakes. Depth ranges for the river reaches are similar to those at Perry Lakes and just as at Perry Lakes, heat budgets in rivers are dominated by radiative fluxes. Sediment conduction accounted for 2.8% of net non advective heat gain and 10.4% of heat loss in the 18 surveys in Table 9.1 completed in 1992 and 1993. These sediment heat flux data display ranges similar to those observed at Perry Lakes (Table 8.3 and Appendix 8.1).

Table 9.1 Mean Daily Bed Conduction (*Qse*) UK Rivers (all values in W m^{-2})

River & Season*	Depth (m)	Gains	Losses	Net
River Piddle Trib (W) 1994	0.50	4.21	-12.94	-8.73
River Piddle Trib (S) 1994		3.10	-51.05	-47.95
River Bere (W) 1994	0.80	4.25	-12.14	-7.89
River Bere (S) 1994		0.47	-0.21	0.26
River Barle Trib 1 (S) 1992	0.12	0.37	-6.06	-5.69
River Barle Trib 2 (S) 1992	0.11	1.20	-15.56	-14.36
River Barle Trib 2 (Sp) 1993	0.12	0.94	-8.41	-7.47
Black Ball Stream (S) 1992	0.10	0.66	-27.25	-26.59
Black Ball Stream (Sp) 1993	0.17	0.44	-7.44	-6.99
Jackmoor Brook (S) 1992	0.19	2.76	-1.33	1.43
Jackmoor Brook (W) 1993	0.17	1.06	-4.66	-3.59
River Creedy Trib (A) 1993	0.15	0.55	-1.59	-1.04
River Pulham (W) 1993	0.51	1.62	-0.17	1.45
River Haddeo (A) 1992	0.16	0.05	-0.84	-0.79
River Haddeo (S) 1993	0.25	1.10	-2.70	-1.60
Iron Mill Stream (S) 1992	0.22	0.10	-1.98	-1.89
Iron Mill Stream (S) 1993	0.26	1.18	-1.32	-0.14
River Haddeo 2 (A) 1992	0.24	2.52	-0.09	2.43
River Haddeo 2 (S) 1993	0.15	0.98	-5.03	-4.05
River Culm (A) 1993	0.48	0.58	-7.96	-7.38
River Culm 2 (S) 1992	0.41	1.52	-12.48	-10.96
River Culm 2 (A) 1993	0.41	0.01	-15.66	-15.65

* S summer, W winter A autumn, Sp spring. Data from Webb & Zhang (1997 & 1999)

In deeper lakes radiative fluxes become less important. Likens & Johnson (1969) obtained thermistor readings to 8m depth in sediments below two small (<1 ha) lakes in Wisconsin. In both the transition from water to sediment was via a gelatinous ooze false bottom. Their purpose was to measure the distribution of sediment heat in small temperate lakes. Tub Lake is of similar size to Perry Lakes (area 0.84 ha) but much deeper (mean 3.6m, maximum 8.0m). Beyond depths of 6m, light levels are extremely low. Stewart's Dark Lake (area 0.69 ha) is similar (mean depth 4.3m, maximum 8.8m). In shallow water, there was a negative steady state gradient indicating a net heat flow into the sediments. In the deeper portions however diurnal variations were completely damped and there was a linear increase of temperature with depth. In Stewart's Dark Lake there was a net positive annual heat flux out of the deep sediments of approximately 0.09 W m^{-2} of which approximately half was solar heat and half geothermal heat.

Long term studies which report only net annual sediment heat flux are deceptive. A lake where annual positive and negative sediment fluxes are equal has a net annual flux of zero. When compared to other studies world wide (Table 9.2), Perry East appears to have an abnormally high negative net sediment flux. This is interpreted to reflect the influence of a large annual net negative advected flux (the lake is predominantly in recharge through storm water and summer top up inputs). In their natural state Perry Lakes probably had net annual sediment heat fluxes closer to the examples in Table 9.2.

Table 9.2 Annual sediment heat flux (*Qse*) for various lakes

Lake	Mean Depth (m)	Sediment Heat Flux (W m^{-2})	Reference
Beloye (USSR)	4.2	-3.32	1
Hula (Israel)	1.7	-1.86	2
Mendota (Wisconsin)	12.1	-2.66	3
Tub (Wisconsin)	3.6	-1.29	4
Stewart's Dark (Wisconsin)	4.3	-0.97	4
Cranberry Pond (New York)	2.9	-2.15	5
Woods(New York)	3.6	-1.49	5
Dart's (New York)	7.1	-0.78	5
Little Simon (New York)	10.0	-0.70	5
Perry East	<1.0	-7.55	This work

References: 1 Rossolimo (1932), 2 Neumann (1953), Birge *et al* (1927) 1,2 & 3 all cited Likens & Johnson (1969), 4 Likens & Johnson (1969), 5 Tsay *et al* (1992). Perry East flux calculated for balance periods 21-50 (January 3 1997 to January 3 1998)

9.3.2 Seasonal Feedback

Likens & Johnson (1969) used a simple model assuming a sinusoidal annual temperature variation in the lake sediments. They found that for these small temperate lakes, maximum feedback of heat from the deepest bottom sediments occurred about 140 days after the seasonal temperature maximum. We also observed a lag at Perry East where the peak occurred around mid August. Ficke (1972) estimated that the negative sediment heat flux in early summer and positive heat flux in early winter for Pretty Lake, Indiana was -8.2 and 14.5 W m^{-2} respectively.

Perry East displays a pronounced seasonal feedback (Figure 9.4). At first sight it might be tempting to equate this with the period of flow through (shown as a shaded bar graph), however it is more likely that this has always occurred even when the lake was flow-through all year. This interpretation is based on the fact that the phenomenon is well documented for temperate Northern Hemisphere lakes most of which are water table lakes in which both groundwater discharge and lake seepage are described. For example at Williams Lake (Minnesota), Sturrock *et al* (1992) describe 'inseeping groundwater' and 'lake water seeping out'. In their thermal balances completed for five summer seasons the daily average sediment term was found always to be negative (*i.e.* a net heat flux from the water column to the sediments). It varied from about -0.05 to -1.03 W m^{-2} over the period April to October. Inclusion of the sediment term in the thermal balance decreased evaporation by up to 7%. It is likely that during winter, after freeze up, Williams Lake probably displays a positive sediment flux but winter data was not collected.

Thermal conditions in shallow lakes and rivers which freeze over winter are highly dependent on heat exchange between the water and sediments. Theoretical calculations (Pivovarov 1973) predict an almost equal balance between heat lost into the sediments over summer and heat returned to the water column from the sediments over winter. Table 9.3 summarises positive winter sediment fluxes for some cold temperate North American lakes.

Table 9.3 Average winter heat flux in temperate lakes

Lake	+ve Flux (W m^{-2})	Reference
Mendota (Wisconsin)	2.9	1
Mendota	3.8	2
Misc Wisconsin Lakes	1.5-1.9	2
Tub (Wisconsin)	1.1	3
Stewart's Dark (Wisconsin)	0.8	3
Pretty (Indiana), November	14.5	4

References: 1 Birge *et al* (1927), Scott (1964), (all cited Likens & Johnson (1969), 3 Likens & Johnson (1969), 4 Ficke (1972).

This also appears to be the case in temperate lakes which don't freeze. Lake Werowrap in the western districts of Victoria (Walker 1973) is a saline lake similar in size and depth to Perry Lakes (mean depth 1.35m, mean surface area 21 ha). Descriptions of lake hydrology suggest that it cycles seasonally between flow through and recharge regimes. In summer groundwater seepage springs around the up gradient shore dry up and the lake probably reverts to recharge status. Monthly sediment heat flux over two years displays a similar pattern to that observed in Perry East (Figure 9.4).

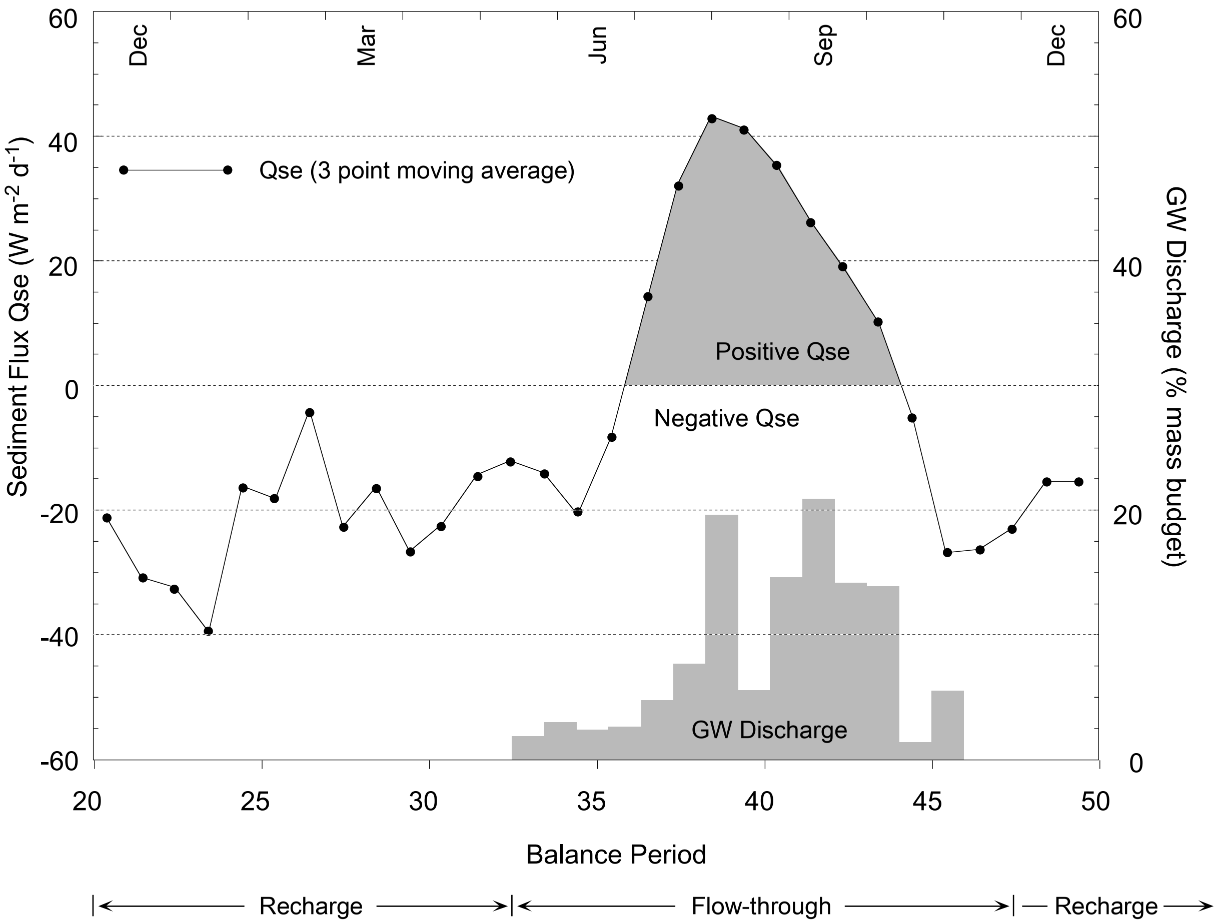
9.3.2 Perry East

Perry East was originally instrumented with the intention that one or a combination of the methods described earlier in this chapter would be employed to measure the sediment heat flux term. The thermal regime in Perry East however is very complex. The water-sediment interface is difficult to define and at any given time this interface displays large areal variations in temperature due to shading from trees and emergent vegetation.

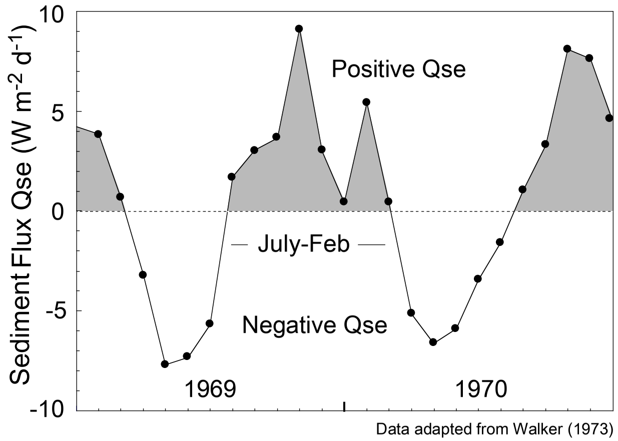
Sediment Heat Flux Qse

Figure 9.4

Perry Lake East Qse by Balance Period



Lake Werowrap Monthly Qse



Positive Qse occurs where there is a net heat flux from the sediment into the water column. Negative Qse occurs where there is a net heat flux from the water column into the sediments

Qse is largely a conductive process driven by the seasonal temperature differential between the water column and the sediments, heat moving from warmer to cooler areas. Qse is influenced by heat advected in groundwater discharge and recharge. Under flow through regimes these advected fluxes occur close to the up and down gradient shores, under recharge regimes water (and heat) is advected over the entire basin.

Perry Lake East, Qse is daily average per balance period,
Lake Werowrap, Qse is daily average per month

This difficulty is illustrated by data from small (<1 ha) temperate lakes in Wisconsin (Table 9.4) where the areal variation of sediment heat flux is defined by water depth. The area of East Lake is also extremely variable and may expand or contract over 100% in less than a day. In shallow lakes anywhere there is a pronounced solar heating of water and sediments. East Lake's shallow depth (mean 0.2 to 0.5m) results in extreme diurnal and seasonal temperature cycles.

As the complexity of the thermal regime became evident it became clear that a holistic method was required rather than the more usual empirical methods where measurements at a single (or several) points are assumed to be representative of the whole. The methodology employed at Perry East was a simple extension of the traditional thermal budget determination of evaporation. If all heat flux terms can be measured, then the residual must be the heat used to evaporate water. In this study, all heat flux terms except the sediment term but including evaporation were measured. The residual is the net sediment heat flux term.

Table 9.4 Spatial distribution of heat flux

Depth (m)	Tub Lake	Stewart's Lake
0-1	-4.05	-3.33
1-2	-2.52	-1.39
2-3	-1.27	-0.76
3-4	-0.53	-0.44
4-5	-0.23	-0.30
5-6	-0.15	-0.19
6-7	-0.11	-0.07
7-8	-0.10	0
8-9	n/a	0

all values in W m^{-2} data from Likens & Johnson, 1969

In their Wisconsin lakes studies Likens & Johnson (1969) estimated that the stored heat contribution from sediments represented approximately 10-12% of the total heat budgets for these lakes. In Chapter 8 we noted that had the sediment flux term been ignored at East Lake and the thermal balance simply used to determine evaporation this would have resulted in a 6.5% over estimate for 1997, equivalent to 89.6mm of evaporation.

Having used the thermal balance to determine sediment heat flux we were left with a huge amount of data from the thermistor strings. While not used for its intended purpose this data allowed the spatial and temporal complexities of the sediment thermal regime to be examined and integrated with the seasonal patterns of lake-aquifer interaction. Section 9.4 examines in detail the instrumentation used to collect the lake and sediment thermal data which is presented in Sections 9.5 and 9.6.

9.4 INSTRUMENTATION

9.4.1 Water temperature loggers

An array of three temperature sensors was constructed to measure surface, mid-level and bottom temperature of the water column (Figure 9.5). Bottom was taken to be clear water immediately above the false bottom. The temperature array was sited in the deepest section of the South Basin (Figure 5.1a). Over the survey period the height of the water column varied from 0.136m (lake stage 2.836m, December 19, 1997) to 0.875m (lake stage 3.575m, September 10, 1997). LM35 temperature sensors were employed. These are completely linear, to better than ±0.25°C (National Semiconductors 1989). Each sensor was calibrated in the laboratory over 0-50°C against a standard laboratory thermometer. Sensors were mounted on a floating frame hinged at the water-sediment interface. Data was captured using three Dataflow 392 single channel loggers mounted in ventilated 40mm PVC enclosures. Point readings were recorded every 10 minutes.

Lake surface temperatures are required for thermal balance (Chapter 8) and some empirical evaporation techniques (Chapter 10). Thermal profiles allowed heat storage (*Qx*) within the lake to be calculated and provide data on stratification and degree of wind mixing. In larger water bodies a number of such profiles are used. This was considered unnecessary due to the small size of East Lake. In such lakes surface temperature is usually measured only in one central location (Anderson 1954 a&b, Harbeck *et al* 1958, Sturrock *et al* 1992).

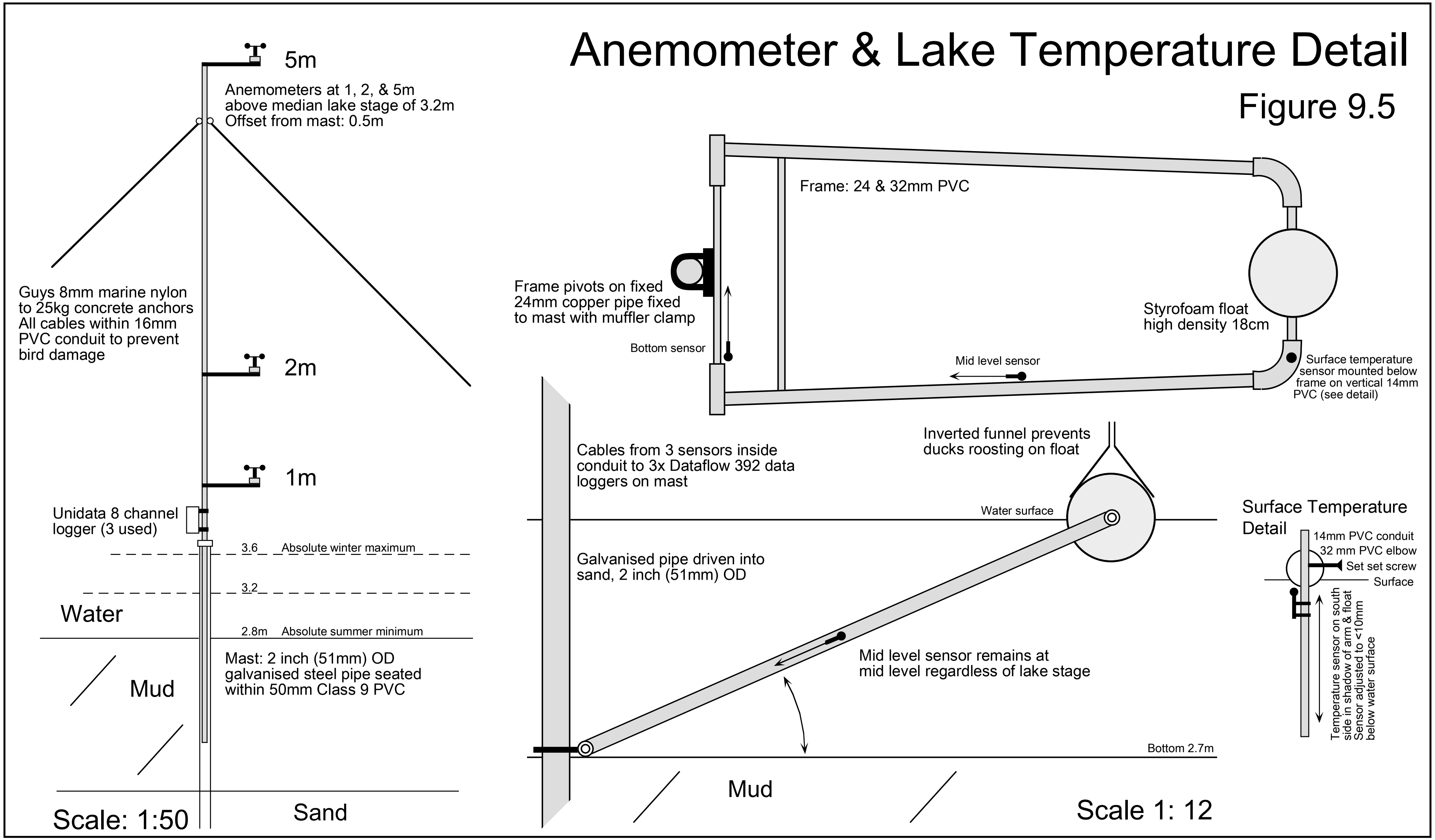
9.4.2 Thermistor strings

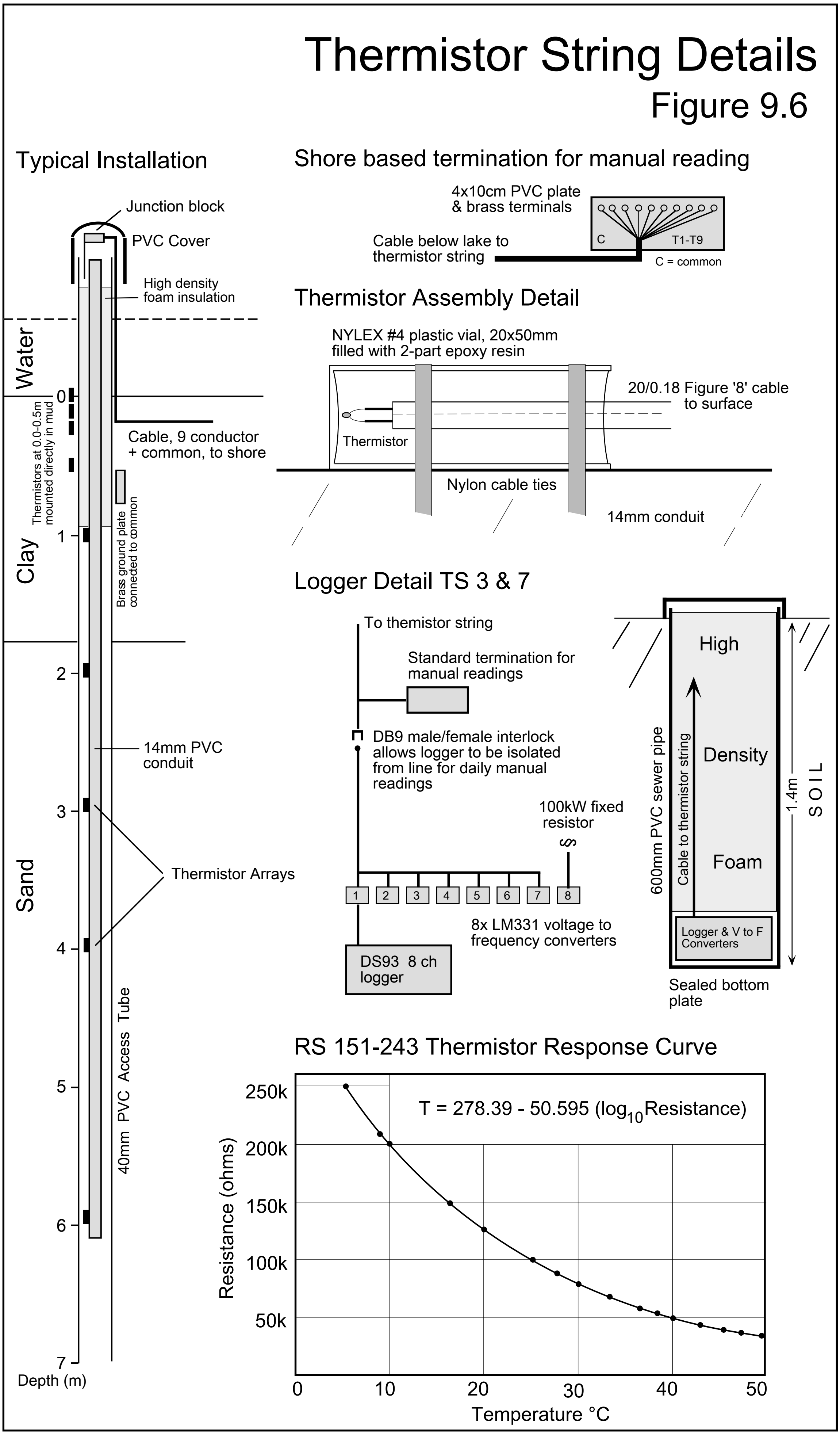
Three profiles of thermistor strings were installed across East Lake (Figures 5.1a and 9.8a). Construction and installation represent modifications of field techniques described by Lapham (1989) and Hunt *et al* (1996). Construction and installation details are summarised in Figure 9.6. The total number of thermistors in a string was limited by the cable (9 conductors plus screened common) connecting the strings to the shore based reading stations. The measured cable resistance was 12 ohms/100m which was considered insignificant in comparison to the operating resistance of the RS 151-243 thermistors (200,000-52,000 ohms between 10-40 °C). Thermistor spacing was varied between different profiles as shown in Table 9.5.

Table 9.5 Thermistor depth below water-sediment interface

Depth (m) refer note	0.0	0.1	0.2	0.3	0.5	1.0	2.0	3.0	4.0	6.0
Ts 2-5 inclusive	X	X	X	X	X	X	X		X	X
Ts 1, TS 6-10 inclusive	X	X	X		X	X	X	X	X	X

Note: Depth is distance below soil surface or water-sediment interface (top of false bottom)


Anemometer & Lake Temperature Detail
Figure 9.5
5m
Anemometers at 1, 2, & 5m
above median lake stage of 3.2m
Offset from mast: 0.5m
Guys 8mm marine nylon
to 25kg concrete anchors
All cables within 16mm
PVC conduit to prevent
bird damage
2m
1m
Unidata 8 channel
logger (3 used)
3.6 Absolute winter maximum
3.2
Water
2.8m Absolute summer minimum
Mast: 2 inch (51mm) OD
galvanised steel pipe seated
within 50mm Class 9 PVC
Mud
Sand
Scale: 1:50
Frame: 24 & 32mm PVC
Frame pivots on fixed
24mm copper pipe fixed
to mast with muffler clamp
Bottom sensor
Mid level sensor
Styrofoam float
high density 18cm
Surface temperature
sensor mounted below
frame on vertical 14mm
PVC (see detail)
Inverted funnel prevents
ducks roosting on float
Cables from 3 sensors inside
conduit to 3x Dataflow 392 data
loggers on mast
Water surface
Galvanised pipe driven into
sand, 2 inch (51mm) OD
Mid level sensor remains at
mid level regardless of lake stage
Bottom 2.7m
Mud
Scale 1: 12
Surface Temperature
Detail
14mm PVC conduit
32 mm PVC elbow
Set set screw
Surface
Temperature sensor on south
side in shadow of arm & float
Sensor adjusted to <10mm
below water surface



Thermistor String Details
Figure 9.6
Typical Installation
Junction block
PVC Cover
High density foam insulation
Water
Clay
Sand
Thermistors at 0.0-0.5m mounted directly in mud
Cable, 9 conductor + common, to shore
Brass ground plate connected to common
14mm PVC conduit
Thermistor Arrays
40mm PVC Access Tube
0
1
2
3
4
5
6
7
Depth (m)
Shore based termination for manual reading
4x10cm PVC plate & brass terminals
Cable below lake to thermistor string
C
T1-T9
C = common
Thermistor Assembly Detail
NYLEX #4 plastic vial, 20x50mm filled with 2-part epoxy resin
20/0.18 Figure '8' cable to surface
Thermistor
Nylon cable ties
14mm conduit
Logger Detail TS 3 & 7
To themistor string
Standard termination for manual readings
DB9 male/female interlock allows logger to be isolated from line for daily manual readings
100kW fixed resistor
1
2
3
4
5
6
7
8
8x LM331 voltage to frequency converters
DS93 8 ch logger
High
Density
Foam
600mm PVC sewer pipe
Cable to thermistor string
1.4m
SOIL
Logger & V to F Converters
Sealed bottom plate
RS 151-243 Thermistor Response Curve
T = 278.39 - 50.595 (log10Resistance)
Resistance (ohms)
250k
200k
150k
100k
50k
0
10
20
30
40
50
Temperature °C

The additional thermistor at 0.3m in Ts 2-Ts 5 was to provide additional detail on diurnal fluctuations and diurnal extinction depths. The additional thermistor at 3.0m in the remainder provided additional long term (seasonal) data.

The thermistor strings were installed within 40mm PVC access tubes installed by sludge pump to 7m depth below the land surface (Ts 1) or water-sediment interface (Ts 2-10). The top of each access tube was 0.5m above anticipated winter maximum lake level. The method is similar to that of Hunt *et al* (1996) with the exception that the access tubes were not sealed at the base. They are essentially piezometers with the water level in each reflecting potentiometric head at 7m depth. Because there is no direct hydraulic connection with the lake however, the water column in the access tube is essentially static and the temperature at any height in the tube is in equilibrium with the aquifer material outside the tube. The alternative was to insert the thermistor string and then withdraw the access tube. Initial experiments suggested this would be near impossible for the 9 lake based sites as the belled and glued access tube joints precluded easy recovery of the tube. The system adopted also allowed any thermistors suffering water leakage to be replaced and ultimate removal of the thermistor strings at the completion of the project.

Initially all thermistors were mounted inside the 40mm access tube. Thermistors in the upper 1m suffered spurious diurnal cycling from changes in lake water temperature. Experiments with thermistors mounted directly in the lake bed mud confirmed that this cycling exhibited different phase and amplitude to that occurring in the mud and was present to about 0.7m depth. An experimental solution was found whereby all thermistors above 1m were inserted directly in the mud approximately 0.4m away from the access tube. Within the access tube, high density foam water pipe insulation was installed from the 1m thermistor to a height exceeding winter maximum lake stage (Figure 9.6). Induced currents were present on the cables from Ts 6 and 7 which manifest themselves as erratically fluctuating resistance readings for all thermistors. These currents originated with high voltage overhead and underground power cables at the sewage pumping station (Figure 5.1a). The currents were damped by grounding the common signal return with a brass plate buried in the mud at the thermistor end of each array (Figure 9.6).

Data from the thermistors (90 in total, 9x10 strings) was collected manually and electronically. Daily manual readings were taken at nominal 08:00hr using a digital multimeter (Altronics DT830B). These readings (individual thermistor resistance in ohms) were entered in spreadsheets and converted to temperature in °C. Data was rounded to 0.1°. Typically one hour was required daily to read all 10 strings. Readings were always completed in the same station order determined by the locations of the cable terminations on shore (Figure 5.1a). These in turn were determined by the locations of suitable hiding places (typically beneath low bushes) to preclude vandalism. The daily manual readings provide high resolution seasonal information useful below the diurnal extinction depth.

Continuous readings were collected from strings 3 and 7 in the Central and South Basins (refer location map Figure 9.8a). Two 8 channel data loggers (Dataflow Systems DS93) were modified to log data from the RS 151-243 thermistors. Principal modification involved the construction of 16 custom voltage to frequency (V to F) converters, built around a precision V to F chip (National Semiconductor LM331). Circuit details (Appendix 9.1) are modified from a typical application design provided by National Semiconductor (National Semiconductors 1989). Each V to F converter was individually calibrated using 37 1% precision resistors varying from 10k to 430k ohms. Each resistor represents a known equivalent thermistor temperature between 76.0° (10k ohm) and -6.6° (430k ohm) and resulted in a unique frequency output for each V to F converter. The frequency data was curve matched to produce a polynomial expression allowing frequency to be converted directly to temperature at better than 0.1° precision. Typically a 5th degree polynomial resulted in a coefficient of determination of $R>0.9999$. Loggers scanned each thermistor string every 2 minutes, recording a mean value every 20 minutes. Seven of the ten thermistors in each string were logged (Table 9.6).

The channel '8' V to F converters on both loggers were hard wired with 100k ohm 1% precision resistors across their inputs. These allowed diurnal temperature induced variations in V to F output frequency to be monitored. During the initial set up of Ts 3 in December 1996, with the logger at ground level, it became evident that diurnal temperature changes in the electronics were producing unacceptably high changes in apparent temperature. The logger was run with 100k ohm fixed loads across all 8 V to F inputs. Thermally induced diurnal signals were evident on all channels, equivalent to 1.4°C drift over 24 hours.

Table 9.6 Logger set up Ts 3 and Ts 7

Thermistor depth (m)	0.0	0.1	0.2	0.3	0.5	1.0	2.0	3.0	4.0	6.0
Ts 3 (Logger S/N 8101)	X	X	X	X		X	X		X	
Ts 7 (Logger S/N 20001)	X	X	X		X	X	X		X	

The thermal drift problem was solved by mounting the loggers in 1.2m deep insulated pits (Figure 9.6). CSIRO tests in Gnangara sand showed that summer diurnal temperature fluctuations are damped to better than 0.1°C at 1.5m depth (J. Smith pers com). Subsequent tests (again with fixed 100k ohm inputs) indicated electronic thermal drift was reduced to better than 0.1°C over 24 hours. Long term absolute drift is not important in this application since we were interested only in the pattern of diurnal variations. The daily manual readings provided long term absolute data.

The RS 151-243 thermistors are negative temperature coefficient types. They are precision curve matched to better than ±0.2°C over 0°-70° range precluding the need for individual calibration. They have a resistance of 100k ohms at 25°C. Calibration data is included in Figure 9.6.

Profile sedimentology is summarised in Table 9.7:

Table 9.7 Thermistor string sedimentology

Thermistor string	Ts 1	Ts 2	Ts 3	Ts 4	Ts 5	Ts 6	Ts 7	Ts 8	Ts 9	Ts 10
Peat	0.0-0.3									
Lake lining clays	0.3-0.4	0.0-0.9	0.0-1.2	0.0-1.2	0.0-0.3	0.0-0.8	0.0-2.4	0.0-0.2	0.0-0.8	0.0-0.4
Sand	0.4-6.0	0.9-6.0	1.2-6.0	1.2-6.0	0.3-6.0	0.8-6.0	2.4-6.0	0.2-6.0	0.8-6.0	0.4-6.0

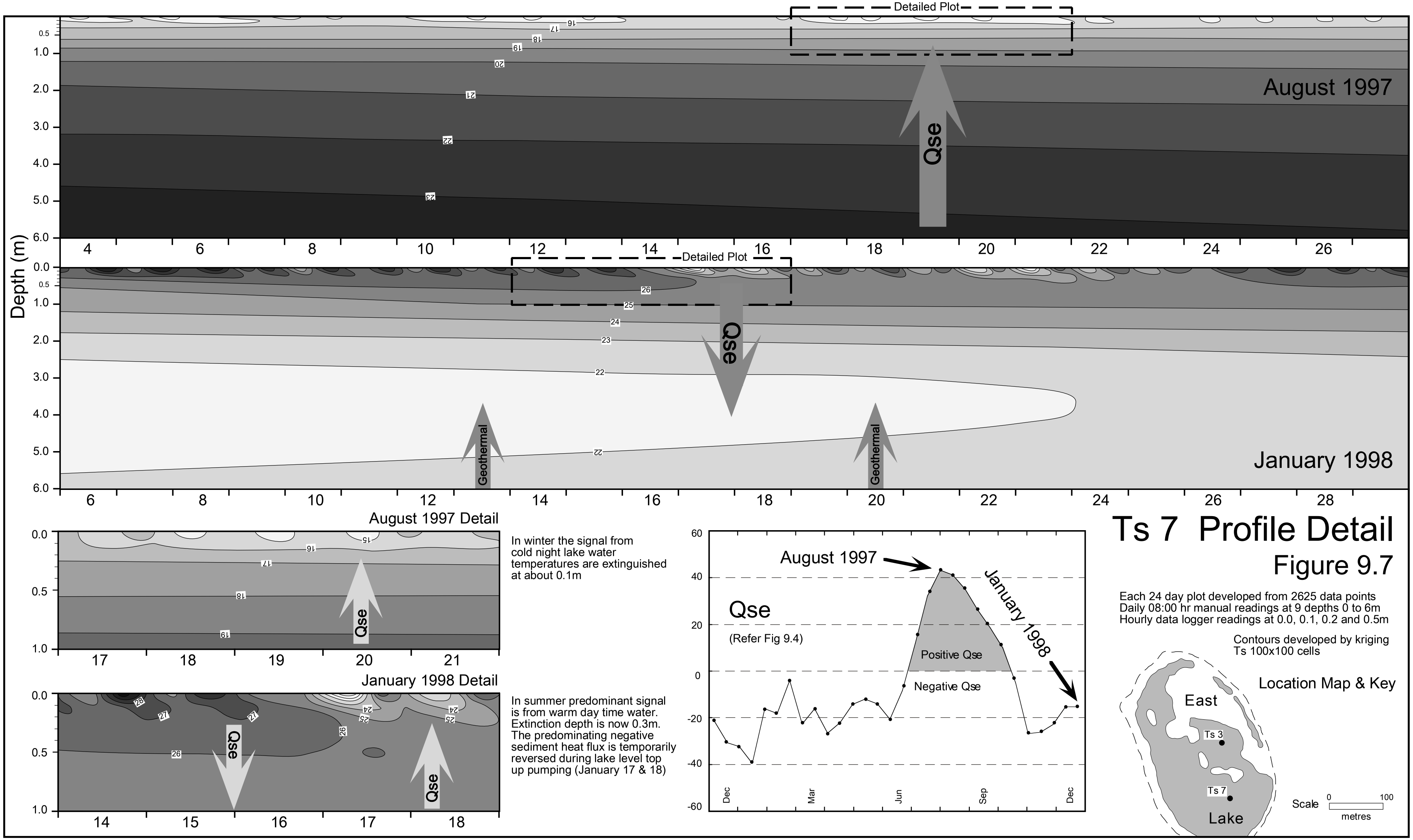
9.4.3 Regional Data

Temperature profiles were measured monthly in the five deep piezometers N1c to N5c. Limited data was also collected for comparative purposes from PL3 and regionally from wells in the UWA Field Station, Bold Park and Jubilee Park (refer location maps within Figure 9.8). An epoxy encapsulated RS 151-243 thermistor was lowered slowly (to prevent vertical mixing) into each well. Temperature was measured at one metre intervals. Each reading took about 1 to 2 minutes, this being the time required for the thermistor to equilibrate.

9.5 THE WATER-SEDIMENT INTERFACE

9.5.1 Diurnal extinction depth

At Perry Lakes there is a continuum between the water column, water saturated unconsolidated gel like 'false bottom' sediments and compact clays. The water column displays strong diurnal temperature cycles of a sin wave nature. This diurnal signal extends into the false bottom where with increasing depth its amplitude decreases and the phase is shifted. The extinction depth is the point where the diurnal signal is completely damped and varies spatially and temporally within the lake basin. Continuous twenty minute thermal data was collected from Ts 3 and Ts 7 (Section 9.5). The data are similar and only Ts 7 data is presented.



Ts 7 Profile Detail

Figure 9.7

During winter, under a positive sediment heat flux, the signal from cold August night time lake water is extinguished at about 0.1m sediment depth (Figure 9.7). During summer a negative sediment heat flux prevails. Hot day time air temperatures heat the lake water and the underlying sediments. This hot diurnal signal is extinguished at about 0.3m. The local flux is reversed during lake level maintenance when large quantities of groundwater at an average temperature of 20.7˚ C are introduced (Figure 9.7 January detail).

Concepts of conductive and advective heat flows are expanded in Section 9.6.

9.6 REGIONAL EFFECTS

Below the extinction depth seasonal patterns predominate. In East Lake these patterns vary spatially across the lake basin reflecting seasonal changes in flow regime. Vertical profiles of groundwater temperature within and close to a flow-through lake reflect heat transported by advection (via groundwater flow) and conduction (Suzuki 1960, Stallman 1965). Hunt *et al* (1996) provide a practical field demonstration of how the groundwater component within a wetland water balance can be estimated from the seasonal depth of heat penetration from the land surface. In their study, the one-dimensional numerical model of Lapham (1989) was used to solve the partial differential equation governing heat flow derived by Stallman (1965). In East Lake this methodology was expanded to include profiles beneath zones of seasonal and permanent inundation however the data were used only to generate temperature profiles for comparison with the flow regimes defined by the water balances.

9.6.1 The regional aquifer

N5 (Figure 9.8d) is an example of a section of the aquifer almost completely insulated from solar effects. Mean soil thickness is about 8.5m. A 'cold front' is conducted down the soil profile over winter, reaching the aquifer by about November. This results in a cold 'tongue' of water (0.1˚C cooler than the surrounding water) which persists until about the following July. The summer 'warm front' also moves down, reaching the aquifer about August the following winter. This warming effect (of up to 0.2˚C) persists until November when the cycle repeats. Data collected regionally from UWA Field Station, Bold Park and Jubilee Park (Figure 9.8f) show similar surface effects superimposed on the regional geothermal gradient which is expressed as a slow warming trend with depth (Davidson 1995).

9.6.2 The aquifer adjacent to wetlands

N1c and N3c are piezometers very close to the original up gradient lake edges. West Lake is now dry much of the year while East Lake has shrunk and is now effectively centred south of N3c (the 'South Basin'). We know that within Perry Lakes Reserve water in the upper part of the unconfined aquifer is of the order of 21˚C. When a flow-through regime is established in either lake over winter warmer water moves upwards from depth and enters the lakes close to the up gradient shores. We observe a hint of this in N3c (Figure 9.8b) in May-June (defined by the field between 20.0 and 20.2˚C) and more weakly in N1c (Figure 9.8f) during June in the field between 20.2 and 20.4˚C.

Both piezometers however display a clear thermal boundary separating surface and possible wetland interaction effects with the deeper regional thermal gradient. In N1c this is defined by the 20.4˚C isotherm which is flat over time at about 13m depth with temperature slowly increasing below that level. Similarly in N3c, the 20.2˚C isotherm is essentially horizontal at about 16m depth. The thermal data corroborates the argument put forward in Chapter 7 that both piezometers were largely outside the zone of surface water-groundwater interaction. The observed seasonal patterns appear to be primarily the effects of conduction from the land surface which is in both N1c and N3c only about 2.5m above the water table. In N1c the thermal patterns over summer can only be surface conduction effects as the adjacent section of West Lake was dry.

Piezometers N2c and N4c on the down gradient shores display patterns which are a combination of surface conduction and flow regime advection effects. Groundwater recharge exceeds groundwater discharge in both lakes. This is particularly so in East Lake where top up maintains persistent summer recharge regimes. In N4c a tongue of warm discharge water (defined by the 22.0˚C isotherm) descends to about 16-17m below the water table (Figure 9.8d). This along with limited data from PL3 provide estimates of the depth of surface water-groundwater interaction. Aquifer thickness is about 36-37m suggesting interaction to about half the aquifer thickness (0.5B). This plume of warm discharge water extends for an unknown distance down gradient. The distance between N4c and PL3 is 64m. There is no evidence of the plume in N5c, 420m distant (Figure 9.8d).

9.6.3 The aquifer beneath the lake basin

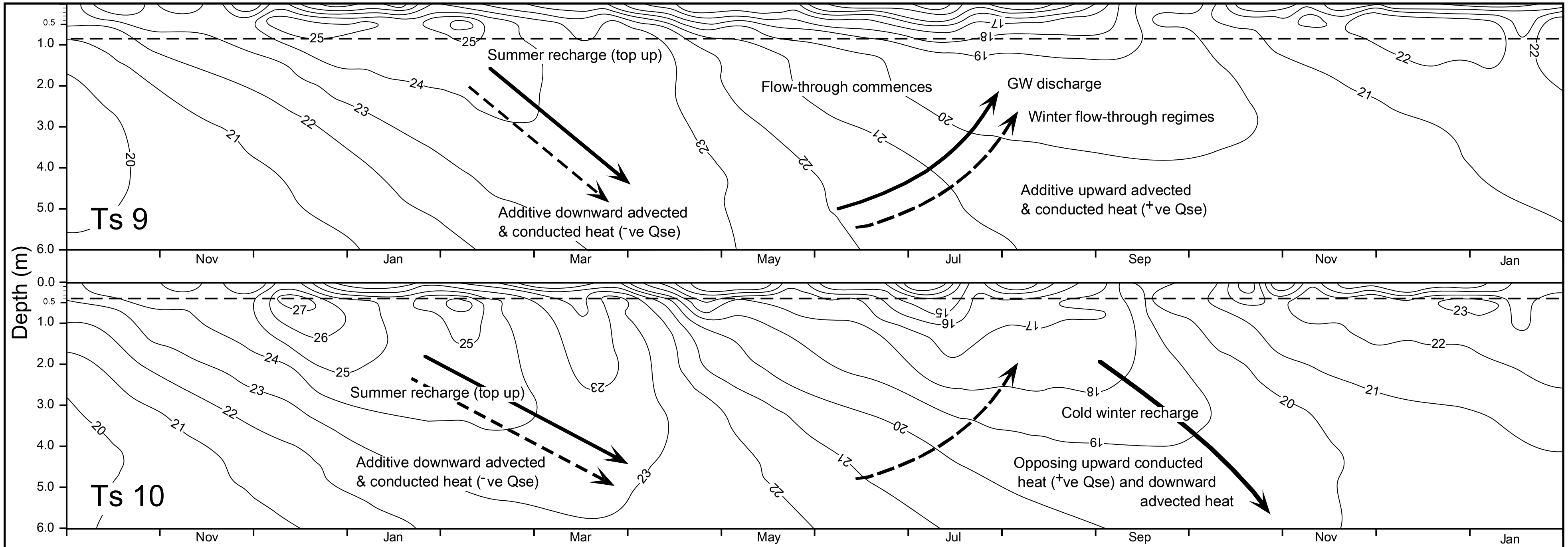
Ten thermistor strings comprising three profiles were installed across the northern, central and southern sections of East Lake (location map, Figure 9.8a). The profiles were aligned parallel to the regional groundwater flow. Each profile displays a similar pattern from east (up gradient) to west (down gradient). The only exception is Ts 1 (Figure 9.8b) which was located close to N3c and also appears to be uninfluenced by surface water-groundwater interaction. In summer the area is dry. Dense *Typha* shades the ground resulting in little heat being conducted into the soil. In winter the area is flooded to about 10cm depth. Winter cold is conducted well below 6m depth.

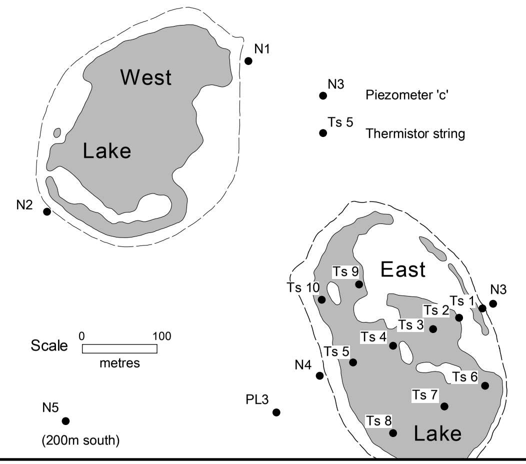
Ts 9, Ts 2 and Ts 6 are located close to the up gradient shore. Their thermal time-depth patterns reflect this in that they are influenced seasonally by both flow-through and recharge regimes. In summer under recharge regimes induced by lake top up, heat is both advected and conducted downwards. In winter under flow-through regimes the process reverses with heat now advected and conducted upwards. This is the positive sediment flux (positive *Qse*) discussed in Section 9.2. The result are annual patterns with much smaller ranges of seasonal change than the equivalent stations (Ts 10, Ts 5 and Ts 8) on the opposite side of the lake.

Ts 10, Ts 5 and Ts 8 are located close to the down gradient shore. In summer heat is again advected and conducted downwards. Compared to holes on the up gradient side, the recharge flux here is much greater. This is particularly evident in Ts 8 (Figure 9.8e) which was adjacent to the South Basin which was almost continually flooded over summer. In comparison Ts 5 and Ts 10 suffered periodic drying out on a weekly basis. In winter the downward advection of lake water continues as the lake reverts to flow-through regimes. Now however there are the opposing forces of upward conduction from sediments warmed the previous summer (positive *Qse*) and the downward advection of cold winter lake water.

Profiles in the centre of the lake Ts 4 and Ts 7 were on or close to the dividing stream line. Heat was advected only in the summer under top up induced recharge regimes. In winter heat transfer was by conduction only, this again being an upward (positive) flux from the sediments.

The thermal profiles also demonstrate the effects of differing management regimes. In early summer 1996, top up commenced October 19. Early top up forced the lake from flow-through to recharge. Warmer summer water advected heat downward. In contrast, in 1997, top up was withheld until December 20 resulting in an entirely different thermal pattern. This is particularly evident in Ts 2 and Ts 5. In Ts 6, a distinct change in slope of the 21° isotherm marks the commencement of top up and transition from a flow-through to recharge regime in December 1997. This effect is demonstrated in Table 9.8 which tabulates the daily rate of temperature rise at fixed depth.





Location Maps & Notes

Advected heat flux
Conducted heat flux

All thermistor strings measure from 0 (sediment-water interface) to 6m depth, thermistor spacing varies, as follows:

Depth	Ts
0.0, 0.1, 0.2, 0.3, 0.5, 1.0, 2.0, 4.0, 6.0	2, 3, 4, 5
0.0, 0.1, 0.2, 0.5, 1.0, 2.0, 3.0, 4.0, 6.0	1, 6, 7, 8, 9, 10

Each thermistor plot developed from 4212 data points (9 daily 08:00 hr readings over 468 days), each piezometer plot developed from about 357 data points (20-21 monthly readings)

Piezometer surveys measure temperature of aquifer from water table to depth of 20-21m (depending on season) measured monthly at metre intervals

All temperatures accurate to better than 0.1°C

Contours develped by kriging: Ts 10x100 cells
Piezometers 20x15 cells

Thermal Profiles in Sediments

Within & Adjacent to East and West Lakes

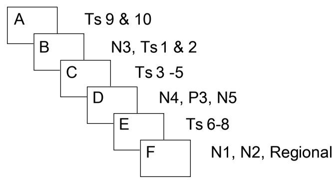
Lake lining (mud) thickness (m)

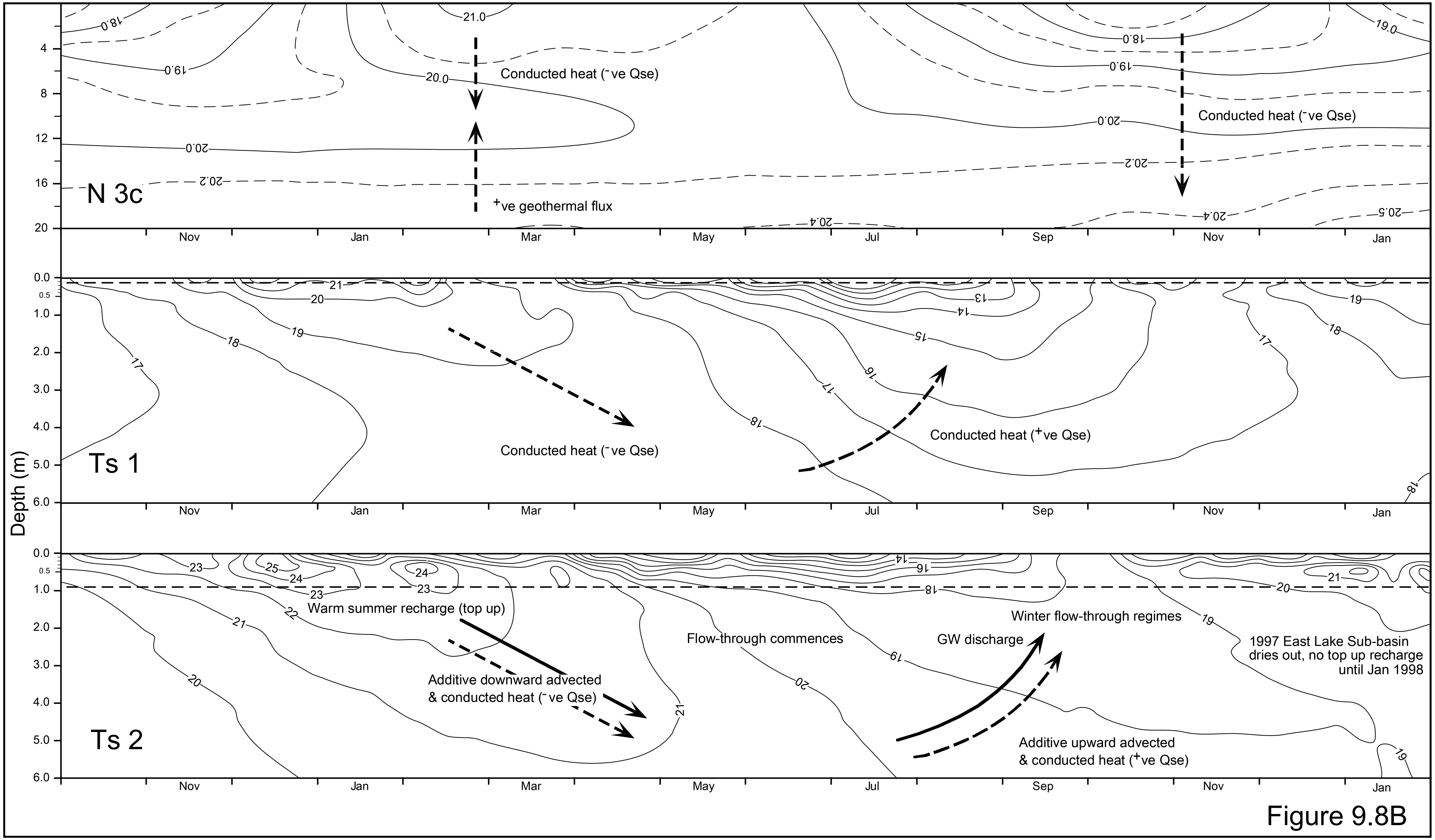
Ts	Mud
1	0.2
2	0.9
3	1.2
4	1.2
5	0.2
6	0.8
7	2.4
8	0.2
9	0.8
10	0.4

Mud-sand contact:

Figure 9.8 A-F

Note: Figure covers 6 sheets





N 3c
Conducted heat (⁻ve Qse)
⁺ve geothermal flux
Conducted heat (⁻ve Qse)
Ts 1
Conducted heat (⁻ve Qse)
Conducted heat (⁺ve Qse)
Ts 2
Warm summer recharge (top up)
Additive downward advected
& conducted heat (⁻ve Qse)
Flow-through commences
GW discharge
Winter flow-through regimes
1997 East Lake Sub-basin
dries out, no top up recharge
until Jan 1998
Additive upward advected
& conducted heat (⁺ve Qse)
Depth (m)
Nov
Jan
Mar
May
Jul
Sep
Nov
Jan

Figure 9.8B

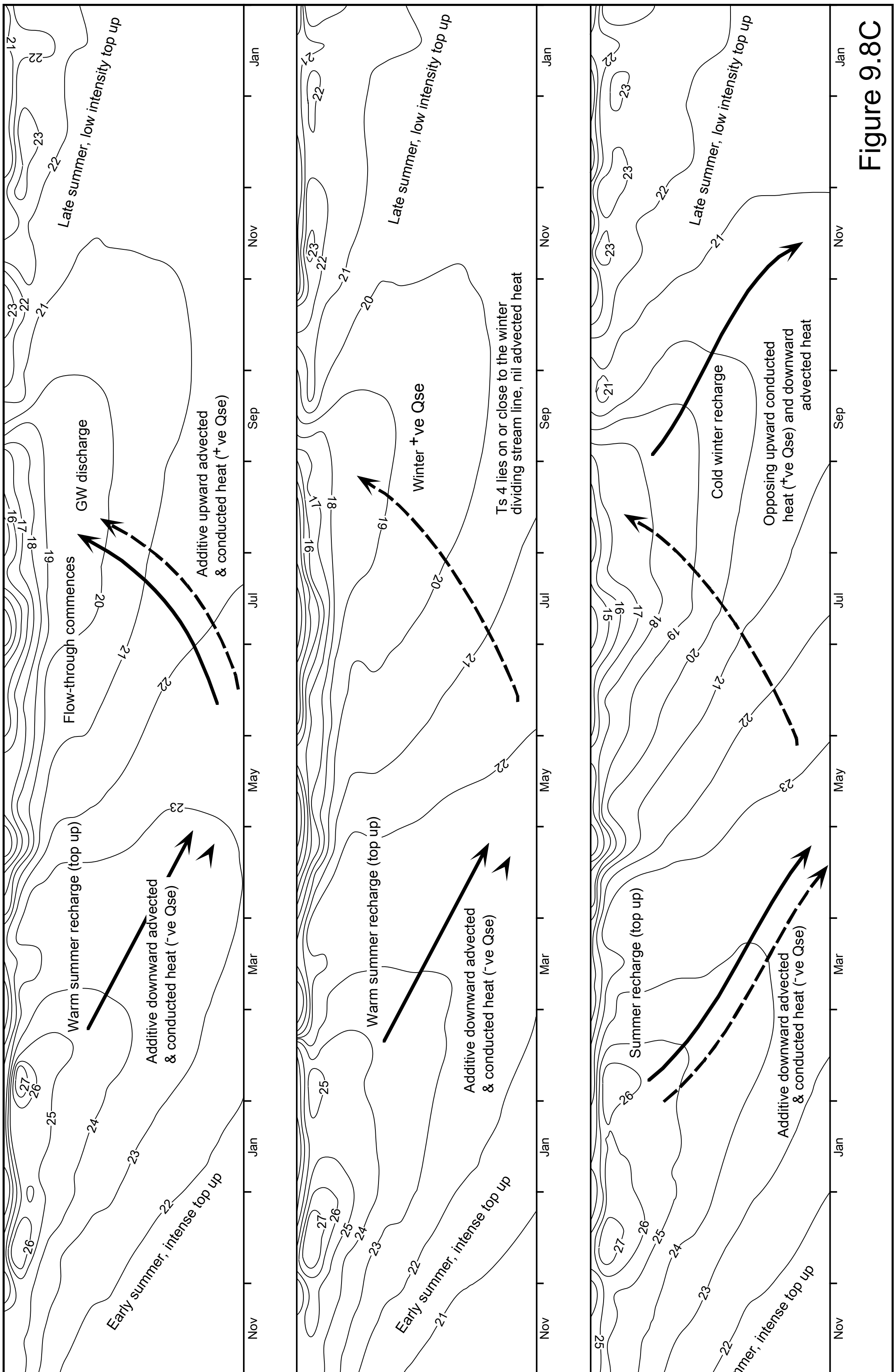

Ts 3
Ts 4
Ts 5
Depth (m)
Nov
Jan
Mar
May
Jul
Sep
Early summer, intense top up
Warm summer recharge (top up)
Additive downward advected & conducted heat (⁻ve Qse)
Flow-through commences
GW discharge
Additive upward advected & conducted heat (⁺ve Qse)
Late summer, low intensity top up
Winter ⁺ve Qse
Ts 4 lies on or close to the winter dividing stream line, nil advected heat
Summer recharge (top up)
Cold winter recharge
Opposing upward conducted heat (⁺ve Qse) and downward advected heat

Figure 9.8C

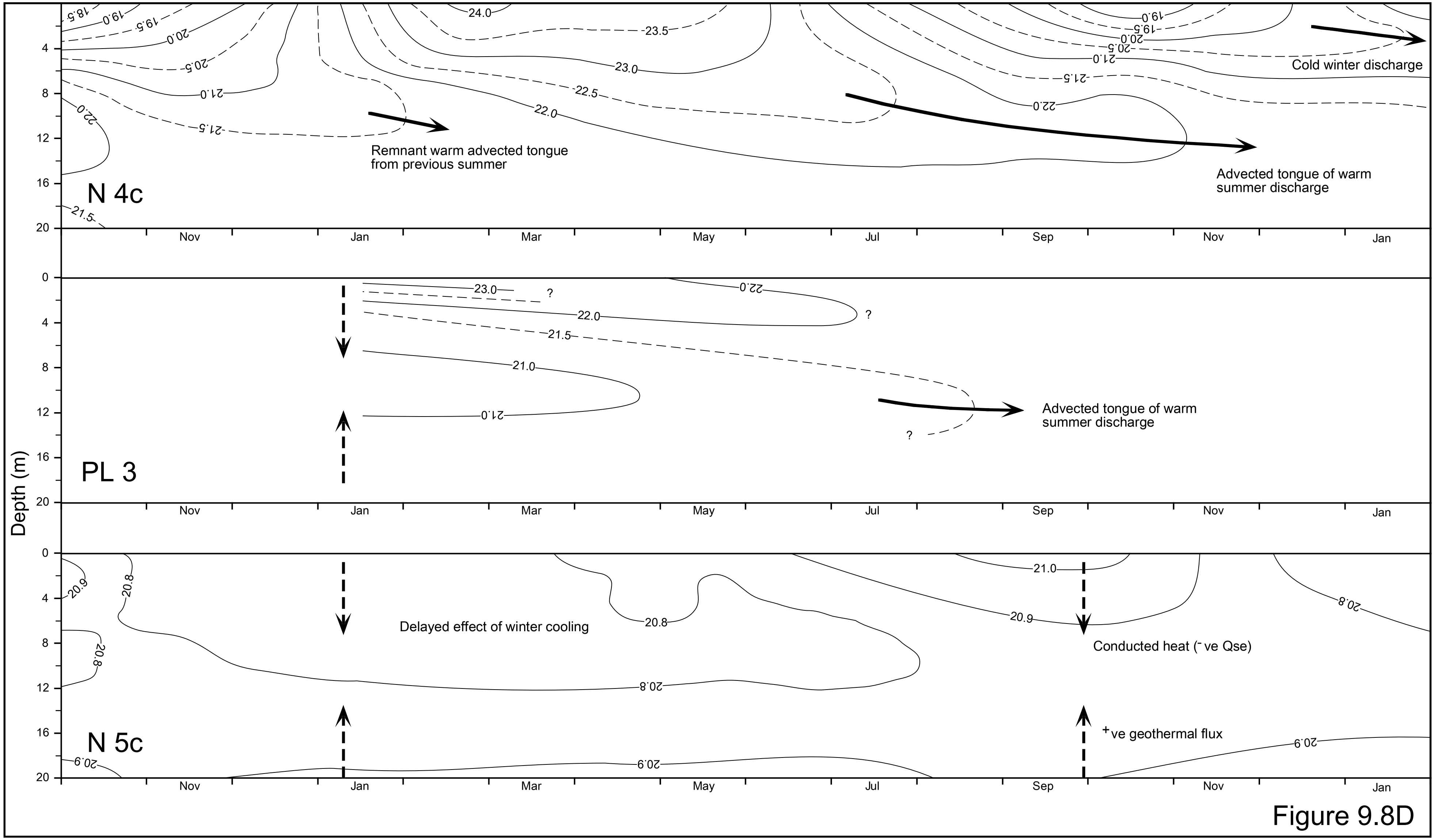

N 4c
PL 3
N 5c
Depth (m)
Remnant warm advected tongue
from previous summer
Advected tongue of warm
summer discharge
Cold winter discharge
Advected tongue of warm
summer discharge
Delayed effect of winter cooling
Conducted heat (- ve Qse)
+ve geothermal flux
Nov
Jan
Mar
May
Jul
Sep
Nov
Jan

Figure 9.8D

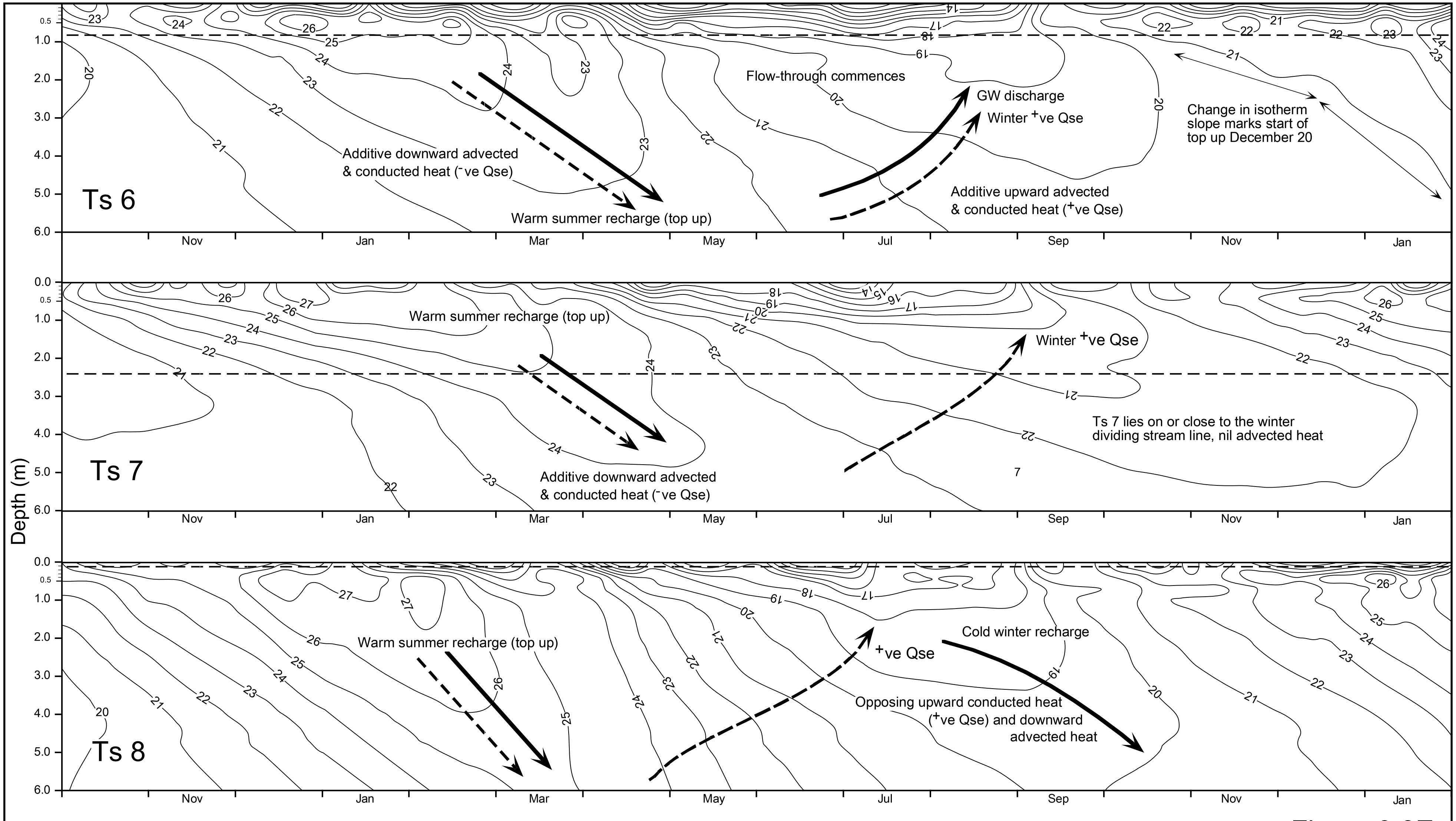
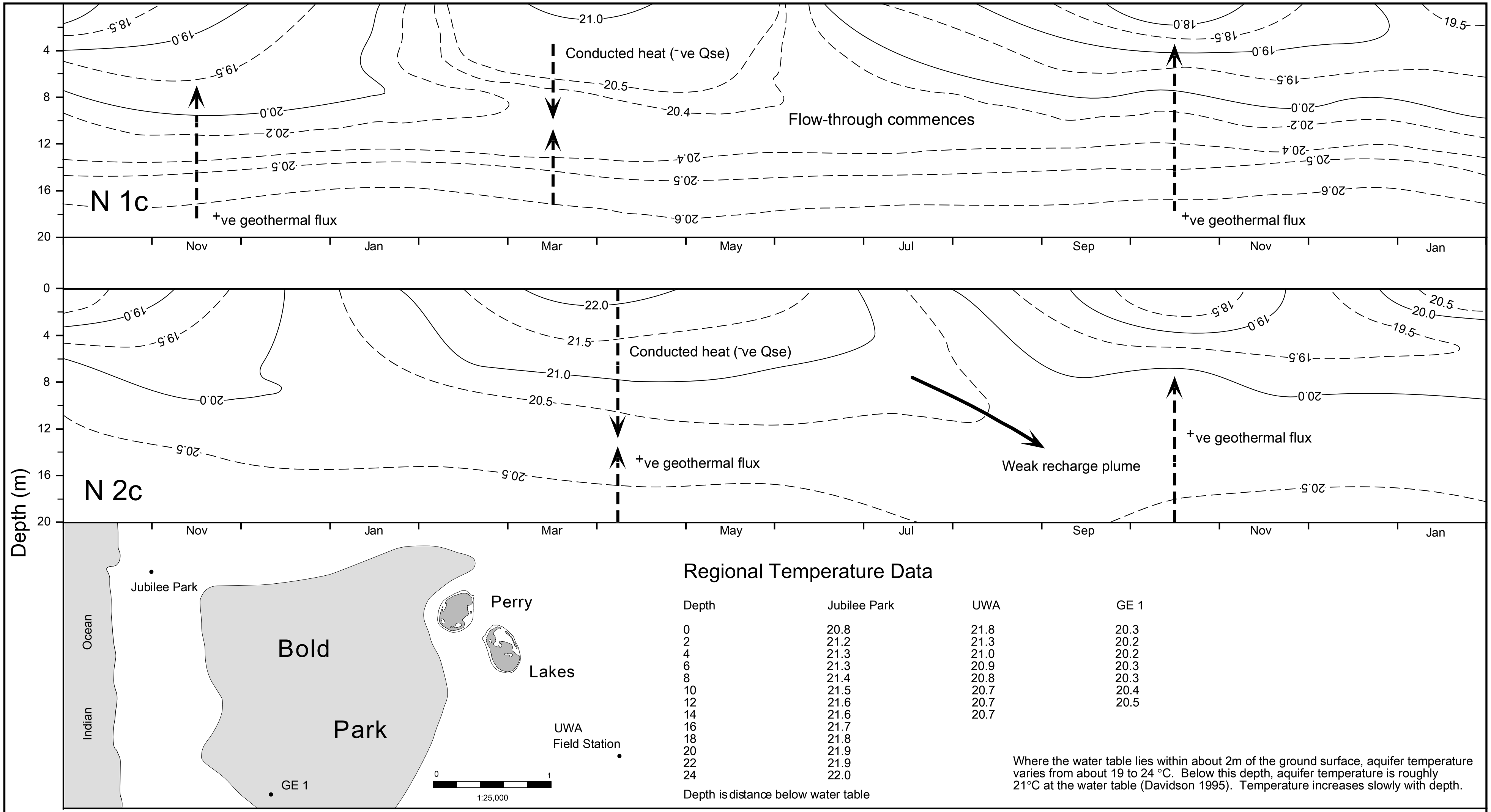

Ts 6
Ts 7
Ts 8
Depth (m)
Nov
Jan
Mar
May
Jul
Sep
Nov
Jan
Flow-through commences
GW discharge
Winter +ve Qse
Change in isotherm slope marks start of top up December 20
Additive downward advected & conducted heat (-ve Qse)
Warm summer recharge (top up)
Additive upward advected & conducted heat (+ve Qse)
Warm summer recharge (top up)
Winter +ve Qse
Ts 7 lies on or close to the winter dividing stream line, nil advected heat
Additive downward advected & conducted heat (-ve Qse)
Warm summer recharge (top up)
Cold winter recharge
+ve Qse
Opposing upward conducted heat (+ve Qse) and downward advected heat

Figure 9.8E



Regional Temperature Data

Depth	Jubilee Park	UWA	GE 1
0	20.8	21.8	20.3
2	21.2	21.3	20.2
4	21.3	21.0	20.2
6	21.3	20.9	20.3
8	21.4	20.8	20.3
10	21.5	20.7	20.4
12	21.6	20.7	20.5
14	21.6	20.7	
16	21.7		
18	21.8		
20	21.9		
22	21.9		
24	22.0		

Depth is distance below water table

Where the water table lies within about 2m of the ground surface, aquifer temperature varies from about 19 to 24 °C. Below this depth, aquifer temperature is roughly 21°C at the water table (Davidson 1995). Temperature increases slowly with depth.

Figure 9.8F

Table 9.8 Sediment temperature rise, early summer

Station	Profile	Daily Rate of Temperature Rise (°C) at Depth of 3m Nov 02-Dec 31 1996	Oct 15-Nov 30 1997*
Ts 9	North	0.032	0.019
Ts 10	"	0.046	0.035
Ts 1	Central	0.015	0.015
Ts 2	"	0.028	0.016
Ts 3	"	0.034	0.021
Ts 4	"	0.042	0.028
Ts 5	"	0.044	0.039
Ts 6	South	0.028	0.017
Ts 7	"	0.011	0.008
Ts 8	"	0.056	0.038

* Ts 9&10 October 15 to November 15

Early top up in 1996 increased the rate of rise at all stations except Ts 1 which further confirmed that it lay outside the influence of surface water-groundwater interaction.

Plotting the ratio of temperature rise in profile end member pairs illustrates the effects of both surface water-groundwater interaction and management regime (Table 9.9). The rate of summer temperature rise is always greater on the down gradient margin of the lake (ratios <1). This is because in summer under flow-through or recharge regimes, there is an additive effect of conducted and advected heat. Early top up and persistent artificially induced summer recharge accentuates this effect. The 1996 ratios are therefore greater than those in 1997 when the natural flow-through regime was allowed to persist into summer.

Table 9.9 Temperature rise ratios, Up & Down gradient pairs

Ratio	Profile	Ratio of Daily Temperature Rise (°C) at Depth of 3m Nov 02-Dec 31 1996	Oct 15-Nov 30 1997
Ts 9:10	North	0.698	0.533
Ts 2:5	Central	0.627	0.401
Ts 6:8	South	0.499	0.455

9.6.4 Long Term Trends

Thermistor string Ts 7 was left in place and was read occasionally up to June 1999. Table 9.10 shows comparisons with temperatures on the same dates in 1996-1998. The time frame of this data is simply too restricted to confirm definite trends. The data suggests however that with persistent low summer levels (augmented by top up), there is a slow increase in thermal input to the aquifer. Very shallow lakes may be effective as heat sources for the aquifer. They heat up quickly in summer and there is direct absorption of thermal energy by the dark mud bottom.

Over this period 1996 to 1999 there continued to be a net decline in the size (and mean depth) of Perry East.

Table 9.10 Thermal trends in sediments at 6m depth, Ts 7, 1996-1999

Day	1996	1997	1998	1999	Trend
January 22		21.73	22.07	21.88	warmer (?)
May 11		23.67	24.10	24.24	"
June 25		23.73	23.91	23.83	"
December 02	21.71	22.12	21.99		"

The data suggests a probable slow warming trend.