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PART 1: EXTERNAL INFLUENCES ON MIXING PROCESSES

by

M T Waters, D A Luketina and J E Ball

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Author (s)

M T Waters, D A Luketina and J E Ball

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The work reported was carried out and published under the direction of the Director of the Water Research Laboratory.

Abstract

Littoral vegetation has long been recognised as an important sink for nutrients and other pollutants from the water column in wetlands and lakes. Mechanisms by which this removal of pollutants occurs in wetlands are poorly understood; however, the transport of substances through a wetland is thought to play a significant role in this removal process.

This report is an assessment of the order of magnitude of transport processes in wetlands subjected to external forcings. These processes were evaluated for a hypothetical "typical" wetland with features representative of existing constructed wetlands. The most important processes causing mixing are found to be wind effects and penetrative convection. Rain may also cause significant mixing; however mean flow velocities, inflow and outflow processes seem to be of little importance. The processes expected to cause significant mixing occur intermittently rather than continuously, so transport of substances within wetlands will be characterised by quiescent periods, during which mixing is limited, interspersed with periods of moderate to severe mixing.

Density stratification is expected to affect mixing processes significantly; building up under the influence of solar radiation according to diurnal and seasonal patterns, and breaking down by the previously mentioned intermittent mixing processes.

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ABSTRACT

Littoral vegetation has long been recognised as an important sink for nutrients and other pollutants from the water column in wetlands and lakes. Mechanisms by which this removal of pollutants occurs in wetlands are still poorly understood; however, the transport of substances through a wetland is thought to play a significant role in this removal process.

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LIST OF SYMBOLS

(In order of appearance)

ρ density
T temperature
μ conductivity
t time
Cp thermal capacity of water at constant pressure
ϕ radiation flux
η absorption coefficient for shortwave radiation
z displacement in the vertical direction (positive upwards from the water surface)
ϕ_s shortwave radiation flux at the water surface
τ_s transmissivity of radiation at the water surface
K_1, K_2 and K_3 constants in the equation for transmission of shortwave radiation through water
ϕ_{lw} longwave radiation flux
ε emissivity of longwave radiation
σ the Stefan-Boltzmann constant
C fraction of cloud cover
ϕ_{atm} long wave radiation flux in the atmosphere
T_{air} air temperature
H_s sensible heat transfer
C_s transfer coefficient for sensible heat
ρ_{air} density of air
cp specific heat of air
Ua wind speed
T_w water temperature
H_l latent heat transfer
C_l transfer coefficient for latent heat
L_w latent heat of evaporation
Q specific humidity
Q_o saturation specific humidity
R_H relative humidity
p air pressure
$\Re_f$ flux Richardson number
b buoyancy flux
m total turbulent mechanical energy
u_f fall velocity of thermals arising due to penetrative convection
α coefficient of thermal expansion of water by volume
g acceleration due to gravity
d depth of water
$\overline{H}$ surface cooling rate
KE turbulent kinetic energy
PE potential energy
ρ_o ambient water density
ε rate of dissipation of turbulent kinetic energy
C_t^f efficiency with which hypolimnion water is entrained into the epilimnion in a penetrative convection event

h	depth of the mixed layer
u_s	water surface velocity
K	dispersion coefficient
u_*	friction velocity
τ	shear stress due to wind
C_D	drag coefficient
$\Re_i$	Richardson number
u	velocity
t_I	time period over which stratification occurs in the absence of mixing
F	fetch length for wind development
$A, B,$ C, D	constants for wave height prediction
l	length scale for mixing processes
v_{turb}	turbulent velocity
E_u	energy of rainfall per unit area
K_{rain}	dispersion due to rainfall
l_m	length scale over which momentum dominates over buoyancy in a turbulent jet
l_{mu}	length scale over which momentum dominates over cross flows in a turbulent jet
D	diameter of inlet to the wetland
v	velocity of water entering the wetland
b_u	width of turbulent jet
x	longitudinal distance from wetland inlet in the downstream direction
S	dimensionless transition parameter for flow regimes due to withdrawal
Q	outflow from the wetland
N	buoyancy frequency
υ	kinematic viscosity of water
L	limiting horizontal scale of outflow
G_r	Grashof number
S_c	Schmidt number
D	diffusion coefficient
δ	half thickness of withdrawal layer
c_2	coefficient for withdrawal layer thickness calculation

1. INTRODUCTION

1.1. Motivation

Over the last two decades or so, there has been an increase in community and political dissatisfaction, both in Australia and internationally, with conventional methods of treating wastewaters (see for example Beder, 1989). Wetlands are increasingly being designed and built, as alternative wastewater treatment systems, and as adjuncts to other treatment systems, existing or proposed. The worldwide interest in their use has reached the stage that the International Association for Water Quality now sponsors biennial conferences on wetland systems for water pollution control. A particular interest in Australia is the ability of wetlands to remove from water bodies the principal nutrients that contribute to eutrophication; these nutrients being Phosphorus and Nitrogen.

One way of classifying wetlands is to consider them as either free water surface or subsurface wetlands. This report is limited to the consideration of free water surface wetlands, which are shallow bodies of water in which emergent, semi-emergent, submergent or combinations of these types of aquatic macrophytes are grown. In such systems, mean flow velocities are negligible compared to velocities induced by external forcings such as wind and other factors; thus it is necessary to examine the processes by which external forcings on the wetland influence the transport of pollutants within the wetland to understand how nutrients are removed from the water column in these systems.

The fundamental aim of this report is to answer the question:

Which external forcings dominate transport processes in the wetland as a whole?

To simplify calculations, the effects of vegetation on transport processes in the wetland have not been considered here; these will be the focus of another report to be published at a later stage.

1.2. Approach

The study of mixing in wetlands has received little attention to date; hence it is necessary to gather data in the systems themselves to determine the nature of the processes. However, in any research programme, it is desirable to know the order of magnitude of the principal parameters before performing field studies. Knowing the magnitudes of the principal parameters, and having an understanding of the conditions affecting these parameters ensures that appropriate data is collected.

1.3. Method

Scaling analyses were performed on a hypothetical wetland using "typical" values of parameters expected on the East Coast of Australia. The results obtained should be applicable to most wetlands in other regions of the world with similar climatic regimes.

Parameter values were obtained from a review of the literature, a survey of Australian constructed wetlands and from data obtained at the West Byron wastewater treatment wetland.

The analyses give orders of magnitudes of parameters for expected operating conditions. The analyses are useful for determining dominant processes, but should be interpreted cautiously and judiciously as there may be large variations from these conditions within a wetland, and very large variations may occur from wetland to wetland. The uncertainty of most terms is likely to be up to several times the estimated values of the parameters themselves.

1.4. Report Structure

The sections of this report cover the following:

1. Introduction of the reasons for performing the study, and an outline of the approach taken.

2. Overview of the general physical properties relevant to constructed wetlands such as physical dimensions, flow rates, velocities and detention times, and the definition of a hypothetical wetland in which parameters are typical of existing wetlands.

3. Analysis of the mechanisms causing density stratification in wetlands and the extent to which stratification would be expected to occur.

4. Analyses of the processes that cause mixing: molecular diffusion; penetrative convection; wind effects; and rain effects.

5. Demonstration of the interaction between stratification and mixing processes by consideration of a destratification event due to a wind mixing event.

6. and 7. Analyses of the influences of inflow and outflow mechanisms on mixing.

8. Conclusions and recommendations for future research.

2. TYPICAL PROPERTIES

A broad range in the sizes, flow rates and mean detention times are found in existing wetlands. In Table 1, typical maximum and minimum dimensions, flow rates and other parameter values are presented for some wetlands; these data were taken from Crites et al (1988), Knight et al (1992) and Crites (1992). The properties of part of the West Byron Wetlands known as Unit 3 are also given; this is a wetland in NSW used for nutrient removal from tertiary treated municipal sewage (Bavor et al, 1992).

TABLE 1:
DIMENSIONS AND OTHER BULK PROPERTIES OF WETLANDS

Typical values are drawn from Crites et al (1988), Knight et al (1992), Crites (1992)

	West Byron Unit 3	Typical Range Minimum	Typical Range Maximum	Units
Average length	45	10	5 000	m
Average width	40	1	500	m
Average depth	0.5	0.1	1.5	m
Surface area	1 800	150	10^7	m^2
Cross sectional area	20	0.5	250	m^2
Volume	900	15	10^6	m^3
Flow rate	3×10^{-3}	5×10^{-5}	10	m^3/s
Nominal detention time	4	1	30	days
Mean through flow velocity	0.13 11	0.001 0.09	0.1 9	mm/s m/d

Very little data is available for analysis of wetland hydrodynamics. To determine what mixing processes are important it is useful to consider a hypothetical "typical" wetland, subjected to conditions that would be expected in the area of interest, coastal New South Wales. A summary of the physical features of such a wetland, and the meteorologic conditions expected to prevail are given in Table 2. Throughout this report, calculations will be based on the typical values assumed in Table 2.

TABLE 2: PARAMETER VALUES FOR A "TYPICAL" WETLAND

<table>
<tr><td colspan="4">Physical Features</td><td>Reference</td></tr>
<tr><td>length
45 m</td><td>width
25 m</td><td>depth
0.5 m</td><td>mean flow rate
3 l/s</td><td>Table 1</td></tr>
<tr><td colspan="4">Influent – Wetland Density Differences</td><td></td></tr>
<tr><td>Temperature difference</td><td>6.5 ^{o}C</td><td colspan="2" rowspan="2">minimal suspended and solute contributions to density</td><td rowspan="2">Table 3</td></tr>
<tr><td>Resulting density difference</td><td>1.3 kgm^{-3}</td></tr>
<tr><td colspan="4">Radiation and Latent Heat Loads</td><td rowspan="2">Fischer et al, 1979</td></tr>
<tr><td colspan="2">Maximum Daytime Shortwave Radiation
400 Wm^{-2}</td><td colspan="2">Latent Heat
4 to 200 Wm^{-2}</td></tr>
<tr><td colspan="4">Temperatures</td><td rowspan="3">Table 3</td></tr>
<tr><td>In air, summer</td><td>30 ^{o}C maximum
20 ^{o}C minimum</td><td>Water, summer</td><td>29 ^{o}C maximum
25 ^{o}C minimum</td></tr>
<tr><td>winter</td><td>20 ^{o}C maximum
11 ^{o}C minimum</td><td>winter</td><td>20 ^{o}C maximum
17 ^{o}C minimum</td></tr>
<tr><td colspan="4">Wind Speeds</td><td rowspan="2">Section 4.4</td></tr>
<tr><td>low, 0 m/s</td><td>medium, 3 m/s</td><td colspan="2">high, 10 m/s</td></tr>
<tr><td colspan="4">Rainfall Rates</td><td rowspan="2">Canterford et al (1987)</td></tr>
<tr><td colspan="4">Maximum of 50 mm/hr</td></tr>
</table>

3. STRATIFICATION

Stratification is the process by which a water body develops density differences with depth. It can result from differences in the physical constituents of the water (such as salt or suspended solids), or from temperature differences. In a water body where less dense water overlies more dense water, its centre of mass is lower than its centre of volume. For this case, the water body has a lower potential energy than if the density was constant with depth.

In this lower energy state more energy is required to provide the same amount of mixing as in a water body where density is uniform with depth; hence stratification markedly affects the transport of constituents due to mixing processes. Temperature gradients as low as 1°C per metre can severely reduce both vertical and horizontal mixing (Imberger and Patterson 1988).

Ambient stratification also significantly affects inflows and outflows. Inflowing water may pass over or under the main body of water depending upon stratification and the degree of mixing present at the inflow; similarly, stratification can drastically affect the extent over which water is "withdrawn" from a water body. These are well-established phenomena in lakes and reservoirs (Imberger and Patterson 1988).

3.1. Causes of Stratification

In a wetland, stratification of the water column could arise due to three factors:

- heat input (observed as temperature differentials);
- solutes; and
- suspended solids

Recorded data from January to July 1992 at the West Byron Wetland are presented from Bavor et al (1992), in Table 3 and Figure 1 for these three factors; also presented are the differences in density that would result from these data. Conductivity was used as a measure of the amount of solutes present in the water column. To assess the density differences due to solutes, a linear relationship between conductivity and density was assumed, with a constant of proportionality of 8×10^{-4} kgm^{-3}/(µS/cm) as recommended by the ASCE Manual in Agricultural Salinity Assessment (Tanji, 1990).

TABLE 3: PHYSICAL PROPERTIES OF WEST BYRON WETLAND INFLUENT

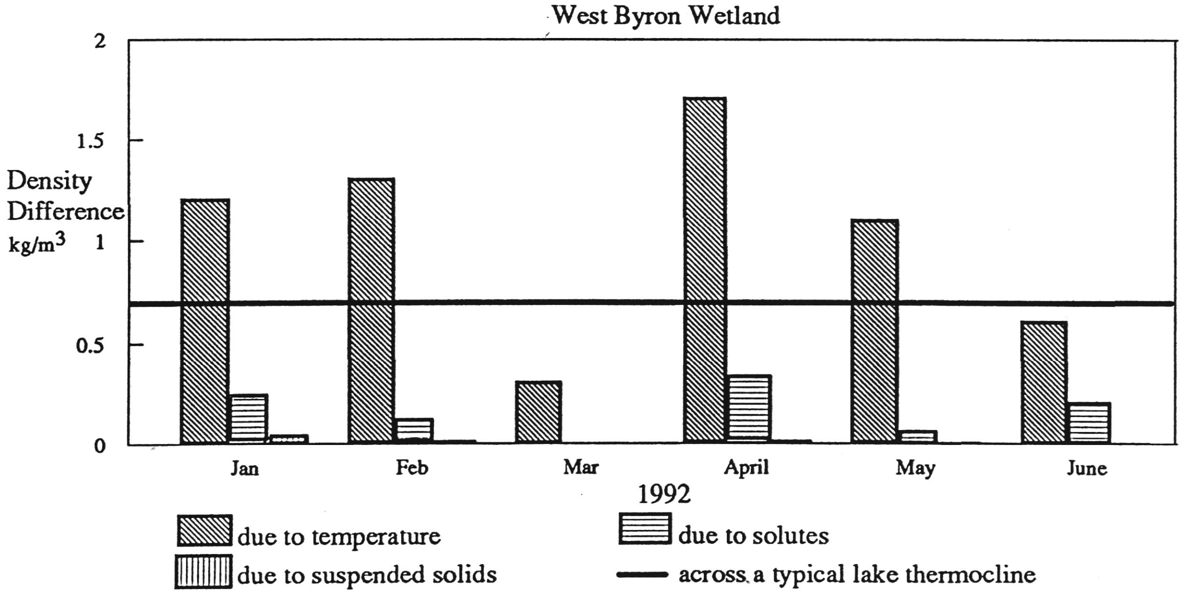
Assuming $\frac{\Delta\rho}{\Delta T} \approx$-0.24 kg/m^3/^{o}C and $\frac{\Delta\rho}{\Delta C} \approx 8 \times 10^{-4}$ kgm^{-3}/(µS/cm)

Month	Temperature T (°C)			$\Delta\rho$ kg/m^3	Conductivity C µS/cm			$\Delta\rho$ kg/m^3	Suspended Solids mg/l		$\Delta\rho$ kg/m^3
	Min	Max	ΔT		Min	Max	ΔC		Min	Max	
Jan	25	30	5.0	1.2	633	905	272	0.2	12	44	0.03
Feb	23.5	29	6.5	1.3	632	766	134	0.1	5	14	0.01
Mar	24.8	26.1	1.3	0.3	Not	Avail			Not	Avail	
April	20	27	7.0	1.7	666	1090	424	0.3	2	12	0.01
May	17.5	22	5.5	1.1	714	789	75	0.06	< 1	4	0.004
June	17	19.5	2.5	0.6	684	875	191	0.2	< 1	4	0.004

FIGURE 1: PHYSICAL PROPERTIES OF INFLUENT

Differences Between Maximum and Minimum Monthly Values of Density in Influent

West Byron Wetland



It is worth comparing the data from Table 3 with other common situations where density differences significantly affect mixing processes. For freshwater and seawater at 20 °C, the density difference can be calculated from Fischer et al (1979) as approximately 26 kgm^{-3} assuming the seawater has a typical salinity of 3.4% by weight. Henderson-Sellers (1984) gives a typical temperature drop across a lake thermocline as 3°C, which is equivalent to

0.7 kgm^{-3} (a thermocline is a distinct temperature gradient existing between the well mixed active upper waters and the more stable lower waters of a lake). By comparison with these other cases, the data reveals that temperature effects on density differences may be high enough to affect mixing processes, while conductivity and suspended solids will have only very minor effects. Thus only the heat-energy balance in the West Byron wetland requires examination to determine its density structure. Note that the residence time of the whole West Byron Wetland is 29 days so it is conceivable that temperature induced density differences as high as 1 to 1.5 kgm^{-3} could arise.

To examine the heat-energy balance, two types of heat transfer processes that may influence stratification must be analysed:

- heating and cooling by short and long wave radiations, which; and
- convective transfer at the water surface by sensible and latent heat transfer. These processes are examined in the sections below.

3.2. Radiation Fluxes

Radiation fluxes result from the absorption or emission of electromagnetic radiation. Two types of radiation are important in the heat balance of a water body:

- Shortwave radiation in the visible spectrum emitted by the sun is largely absorbed in the top 100 to 200 mm of the water column; and
- infra-red radiation. This occurs as longwave radiation emitted by clouds in the atmosphere and as black-body radiation emitted from the water surface.

3.2.1. Shortwave Radiation

In the absence of mixing processes the change in temperature of a water body due to shortwave radiation from the sun penetrating the water column can be expressed as:

$$\frac{\partial T}{\partial t} = \frac{1}{\rho C_p} \frac{\partial \phi}{\partial z} \qquad 1$$

where $\frac{\partial T}{\partial t}$ is the time rate of change of temperature (°C/s), C_p is the thermal capacity of water at constant pressure (J/kg/°C), $\frac{\partial \phi}{\partial z}$ is the vertical gradient of the shortwave radiative flux that penetrates into the water body (Wm^{-2}/m) and ρ is the density (kgm^{-3}).

From Henderson-Sellers (1984), the dependence of ϕ on depth is commonly expressed in exponential form; this gives the mathematical expression known as Beer's Law:

$$\phi(z) = \phi_s e^{-\eta z} \qquad 2$$

where ϕ_s is the radiative flux that penetrates the surface and η is a constant. However, very close to the surface (less than 0.5 m) this expression underestimates the amount of radiation absorbed (Henderson-Sellers 1984). Since wetland depths are typically of the order of 0.5 m, this expression is inappropriate. Henderson-Sellers (1984) presents an empirical expression that is valid at shallow depths as:

$$\phi(z) = \phi_s \tau_s e^{-K_1 z}[1 - K_2 \tan^{-1}(K_3 z)] \qquad 3$$

In Equation 3 the values of the *K* parameters are constants for a particular water body. Typical ranges from Jerlov (1976) for the *K* parameters, from clearest conditions to most turbid, are:

- K_1, 0.04 to 0.6 m^{-1};
- K_2, 0.4 to 0.5 (dimensionless); and
- K_3, 4.4 to 3.6 m^{-1}.

τ_s is the surface transmissivity of the lake which is determined by the light's angle of incidence on the water surface, the optical refractive index of the water and the polarisation of the light. When light is incident on the water surface at right angles the surface transmissivity is 98%, but this decreases rapidly for non-normal radiation.

During daylight hours on a typical sunny day in summer, a typical average shortwave flux value is $\phi_s = 400$ Wm^{-2} (Fischer et al 1979). Transmissivity at the water surface peaks at 98% (Hecht 1987), and typical K values are $K_1 = 0.2$ m^{-1}, $K_2 = 0.4$ and $K_3 = 4$ m^{-1} (Henderson Sellers 1984). By substituting Equation 3 into Equation 1, with the constants evaluated using the values presented above, the following equation is obtained:

$$\frac{\partial T}{\partial t} = \frac{1}{\rho C_p}\{390e^{-0.2z}\{0.2[0.4\tan^{-1}(4z)-1]-\tfrac{1.6}{1+16z^2}\}\} \qquad 4$$

where $\rho \approx 1000$ kg/m^3 and $C_p \approx 4.2x10^3$ J/kg/K are typical values for any freshwater body.

To evaluate the change in density due to the changes in temperature, the rate of change of density with temperature, between 20^{o}C and 25^{o}C, can be approximated as:

$$\frac{\partial \rho}{\partial t} \approx -0.24 \frac{\partial T}{\partial t} \quad 5$$

Without detailed meteorologic records from a specific site, it is difficult to determine over what period of time the conditions in Equation 4 will remain constant. Certainly these conditions cannot prevail for longer than the number of daylight hours (say 12 hours). Assuming that conditions remain consistent over a 5 hour period within this day, Equations 4 and 5 predict the changes in temperature and density shown in Table 4 and Figure 2. Thus for typical sunny, quiescent conditions on a cloud free summer day, an average temperature gradient as high as 4.7°C/m may be established. A stratification of this magnitude is sufficiently severe to affect mixing processes.

TABLE 4:
TEMPERATURE AND DENSITY DEPENDENCE ON DEPTH

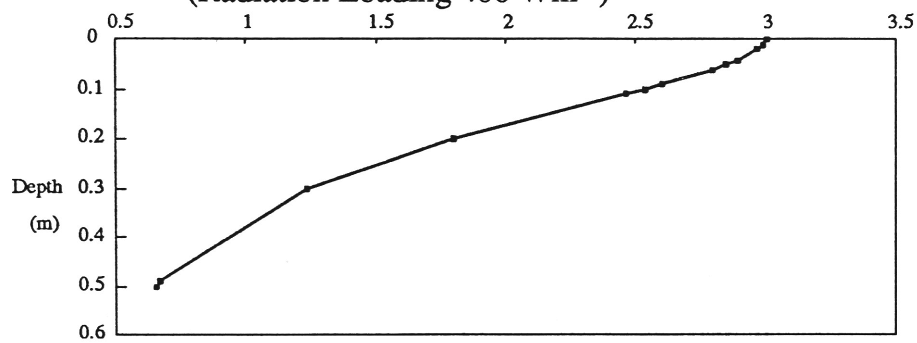
Resulting from an incident shortwave radiation load of 400 W/m^2 for 5 hours

Depth (m)	Change in Temperature (°C)	Change in Density (kg/m^3)
0.0	3.0	–0.72
0.01	3.0	–0.72
0.02	3.0	–0.71
0.05	2.9	–0.70
0.1	2.5	–0.68
0.2	1.8	–0.67
0.5	0.65	–0.67

FIGURE 2 STRATIFICATION PROFILE

Change in Temperature (°C) after 5 hours,

(Radiation Loading 400 Wm^{-2})



3.2.2. Longwave and Blackbody Radiation

All materials with a temperature above absolute zero emit longwave radiation according to the Stefan-Boltzmann law for blackbody radiation, which can be expressed as:

$$\phi_{lw} = \varepsilon \sigma T^4 \qquad 6$$

where ε is the emissivity (0.97 for water), σ is the Stefan-Boltzmann constant which is 5.6699×10^{-8} W m^{-2} K^{-4}, and T is temperature on the Kelvin scale. In wetlands, two types of longwave radiation will be significant:

- longwave radiation emitted by water that is transferred to the atmosphere, and
- longwave radiation emitted by the atmosphere transferred to the water.

Both of these are surface phenomenon, as the coefficient of radiative absorption for water in the infrared region of the spectrum is very high, ranging from 200 to 3000 cm^{-1} over different wavelengths (Perry and Chilton, 1973). Such high absorption coefficients ensure that:

- all long wave radiation impinging on the water surface is absorbed within millimetres of the surface; and
- when long wave radiation is emitted by water, surrounding molecules reabsorb the radiation unless the radiating molecule is very close to the surface.

Longwave radiation emitted from the water surface will be referred to from here as "blackbody radiation". Expected values of blackbody radiation as calculated by the Stefan-Boltzmann Law are given in Table 5.

TABLE 5:
EXPECTED EXTREME VALUES OF BLACKBODY RADIATION EMITTED FROM THE WETLAND SURFACE, INCOMING LONGWAVE RADIATION FROM THE ATMOSPHERE AND SENSIBLE HEAT TRANSFER

All in Wm^{-2}, positive values indicate heat transfer from the water to the atmosphere, sensible heat was calculated for a wind speed of 3 m/s

Conditions	T_{air} °C	T_w °C	Blackbody Radiation ϕ_{bb}	Atmospheric Longwave Radiation ϕ_{atm}	Surface Heat Transfer H_s	Nett Heat Transfer $\phi_{bb}+\phi_{atm}+H_s$
Summer - Day	30	25	434	-407	-27	0
Summer - Night	20	29	457	-333	49	173
Winter - Day	20	17	389	-333	-16	40
Winter - Night	11	20	405	-276	49	178

Longwave radiation from the atmosphere is another source of incoming radiation that impinges on the water surface; this latter form of radiation will be referred to from here as "longwave radiation".

As discussed by Fischer et al (1979), the longwave radiation from the atmosphere, calculated in accordance with the Tennessee Valley Authority (TVA) method, is given by:

$$\phi_{atm} = -5.18 \times 10^{-13}(1+0.17C^2)(273+T_{air})^6 \qquad 7$$

where C is the fraction of cloud cover and T_{air} is the air temperature 10 m above the water surface. The minus sign denotes that longwave radiation transfers heat into the wetland. Over the length of a typical summer day, a representative value of C is 0.3. Representative values of ϕ_{atm} are given in Table 5.

The effects of longwave and blackbody radiation on the thermal structure of the water body are considered in combination with sensible heat transfers in Section 3.3.1.

3.3. Surface Fluxes

Surface heat transfer processes involve the exchange of heat at the air-water boundary. There are two processes of this type; these are the sensible and latent heats of transfer.

3.3.1. Sensible Heat Transfer

Sensible heat transfer is a two phase turbulent transfer process between the water body and atmosphere. It arises from the temperature difference between the water and atmosphere and may be a heat flux into or out of the water body. From Fischer et al (1979), values of sensible heat transfer are calculated as:

$$H_s = C_s \rho_{air} c_p U_a (T_{air} - T_w) \qquad 8$$

where H_s is the sensible heat transfer (defined to be positive when heat leaves the water surface), C_s is the dimensionless transfer coefficient for sensible heat, ρ_{air} is the air density (kgm^{-3}), c_p is the specific heat of air (J/kg/°C), U_a is the wind speed in m/s, 10 m above the water surface , T_{air} is the air temperature (°C) and T_w is the water temperature (°C). In Table 5 maximum positive and negative values are presented for sensible heat transfer under typical conditions, where C_s is 1.45×10^{-3}, ρ_{air} is 1.2 kgm^{-3}, c_p is 1012 J/kg/°C and with an assumed wind speed, U_a is 3 m/s.

Compared to typical incoming solar radiation fluxes (up to 400 Wm^{-2}), the combined effects of longwave radiation, blackbody radiation and sensible heat transfer are only small during the day, but can give rise to a major efflux of heat from the wetland through its surface at night. Thus these processes do not have a significant effect on the stratification process; however, as will be seen in Section 4, they are significant in driving destratification and mixing through penetrative convection.

3.3.2. Latent Heat Transfer

Latent heat transfer arises from the cooling effect of evaporation on the water body. An expression for latent heat transfer is given in Fischer et al (1979) as:

$$H_l = C_l \rho_{air} L_w U_a (Q_o - Q) \qquad 9$$

where C_l is the dimensionless transfer coefficient for latent heat and L_w is the latent heat of evaporation of water (J/kg). Q is the Specific Humidity 10 m above the water surface; that is, the ratio of the partial pressure of atmospheric water vapour to the total air pressure, corrected for the density difference between water vapour and air. Q_o is the saturation specific humidity. Both Q and Q_o depend on T_{air}, according to Chow et al (1989) as:

$$Q = 380\frac{R_H}{p}e^{\frac{17.27T_{air}}{237+T_{air}}} \tag{10}$$

where p is air pressure, R_H is the relative humidity and T_{air} is the air temperature in °C. Note that Q_o is obtained by taking R_H as 100%.

Fischer et al (1979) found in a study performed on a typical summer day in Australian conditions, latent heat varied between 4 and 200 Wm^{-2} with peak values occurring in the late afternoon and early evening. As for the sensible heat case, latent heat is significantly less than shortwave radiation during the day and hence has only a secondary effect on the development of stratification; however, it can be significant in destratification and mixing due to penetrative convection (see Section 4).

3.4. Conclusions

Stratification will play a major role in wetland mixing processes, changes in temperature of up to 3°C over a typical wetland depth of 0.5 m within 5 hours are possible. Such a time scale is significantly shorter than typical residence times, so that water flowing through the wetland has more than enough time to stratify before reaching the exit. Shortwave radiation dominates the development of stratification since longwave radiation into and out of the wetland are approximately equal, and surface heat transfers are generally an order of magnitude smaller than the radiation effects.

4. MECHANISMS FOR MIXING AND DESTRATIFICATION

Many previous investigations of constructed wetlands for wastewater treatment have treated them according to chemical reactor theory (Kadlec, 1993). Such studies have shown that wetlands obey neither plug flow, nor completely mixed behaviour; however, calibrated networks of plug flow and completely mixed reactors, coupled with first order decay models can predict removal rates of chemical constituents from a wetland as a whole. These networks describe the overall behaviour of the wetland well, but they are incapable of providing details of hydrodynamic behaviour within the wetland. As such they cannot be used to determine rates at which constituents become available to, or are removed from sites within the wetland.

The approach preferred here is to identify processes that cause mixing, and quantify their effects on mixing within the wetland by the rates of dispersion to which they give rise. Dispersion is most easily understood as the spreading rate of a tracer cloud. This section evaluates the orders of magnitude of heat dispersivities (sometimes referred to as eddy diffusivities for heat or heat dispersions) due to various mechanisms. With the exception of molecular diffusion, all of the mechanisms described below cause dispersion by increasing the amount of turbulence present in the water column. The focus is on the dispersion of heat, to maintain continuity with the previous sections on stratification and destratification and because the nature of dispersion due to turbulence is intimately connected to the degree of stratification present (Turner, 1973). If the energy associated with the stratification is high enough, then turbulent mixing can be severely suppressed; however if the energy associated with turbulent mixing is sufficiently large, the stratification may be completely overcome.

Two points require clarification before continuing are:

- Under turbulent conditions dispersion of heat and other flow constituents are of the same order of magnitude, so that values derived here for heat dispersion provide an appropriate estimate for other conservative constituents in the flow.

- The term dispersion is used throughout this report to describe the degree of mixing in a wetland. What is meant by the term is the quantity that Fischer et al (1979) call the "turbulent equivalents to molecular diffusion."

4.1. Molecular Diffusion

In the absence of turbulent mechanisms for mixing, molecular diffusion will still take place. Perry and Chilton (1973) list molecular diffusivities of various dilute solutes in water at 25°C. Values are between 0.48×10^{-9} for lactose and 5.9×10^{-9} m^2/s for Hydrogen, the value for ammonia is 2×10^{-9} m^2/s. The diffusivity of heat is significantly higher than that of dilute solutes, being 1.4×10^{-7} m^2/s.

4.2. Turbulent Mixing Processes

In most open water systems where water quality is important, mixing processes are generally turbulent. Turbulence is characterised by:

1. the transfer of energy from the mean flow to the turbulence at the turbulent macroscale;
2. the transfer of mechanical to heat energy at the microscale of the turbulence due to the viscosity of the liquid; and
3. eddy motions between the macroscale and the microscale, which transfer energy between these scales and give rise to mixing.

At the macroscale, the sources of turbulence that are expected to be important in wetlands are:

- penetrative convection;
- mixing due to wind shear; and
- rain effects.

These will be considered in Sections 4.4 to 4.7. Before examining these processes it is desirable to examine the efficiency with which turbulent processes generally give rise to mixing.

4.3. Efficiency of Turbulent Mixing Processes

Ivey and Imberger (1991) defined the flux Richardson Number as the ratio of the energy associated with buoyancy (the buoyancy flux) b, to the total mechanical energy associated with turbulence m, as $\Re_f = \frac{b}{m}$. In processes where turbulence gives rise to a positive buoyancy flux, such as mixing due to shear, $\Re_f$ represents the efficiency of mixing. In processes where a negative buoyancy flux gives rise to turbulence, such as penetrative convection, $\Re_f^{-1}$ represents the efficiency of mixing. They found that generally the energy efficiency of mixing in these turbulent processes is of the order of 10%.

When turbulent mixing processes are described in terms of dispersion and velocity scales, it is necessary to incorporate the efficiency considerations from above. However, dispersion and velocity are proportional to the square root of the energy, so when dealing with velocity and dispersion scales, the efficiency of these processes will be proportional to the square root of the energy efficiency, that is, approximately 30%.

4.4. Penetrative Convection

Penetrative convection occurs when heat transfer from a water body to the atmosphere cools the surface water causing it to become more dense than the water below it, leading to unstable stratification. Hence the surface water descends through the water column, causing vertical mixing. Such conditions may occur soon after nightfall following a warm summer day, and may also occur due to changing climatic conditions, such as the advance of cold fronts.

From Fischer et al (1979), the free-fall velocity of thermals resulting from penetrative convection in an unstratified water body is:

$$u_f = \{\frac{\alpha g d \overline{H}}{C_p \rho}\}^{\frac{1}{3}} \qquad 11$$

where α = coefficient of thermal expansion of water by volume $\approx 2\text{x}10^{-4}$ °C^{-1}

d = water column depth (m);

$\overline{H}$ = surface cooling rate (Wm^{-2});

C_p = heat capacity of water at constant pressure $\approx$ 4179 J/kg/°C

In Section 3, it was found that surface heat transfers from water to the atmosphere consists of sensible heat, which is typically less than 50 Wm^{-2}, and latent heat, which typically ranges from 4 to 200 Wm^{-2}. Taking $\overline{H}$ = 50 Wm^{-2}, as a conservatively small estimate of the rate at which the wetland loses heat gives $u_f \approx 0.002$ m/s. Under these cooling conditions, for a typical wetland in an unstratified state, a plume due to penetrative cooling will reach the bottom of the typical wetland within 200 seconds (3 minutes). It can be seen that under such conditions, which are by no means abnormal, mixing in the wetland occurs fairly quickly.

In a stratified water body, calculation of the rate at which penetrative convection proceeds requires the computation of the mixed layer development. This is done by an energy balance on the layer; a method for this computation is presented by Fischer et al (1979). This energy balance requires that the rate of change of turbulent kinetic energy (KE), the rate of change of potential energy (PE) due to the mixed layer deepening into the unmixed zone and the rate of dissipation of turbulent kinetic energy ε sum to zero; that is:

$$\frac{dKE}{dt} + \frac{dPE}{dt} + \varepsilon = 0 \qquad 12$$

Fischer et al (1979) show that this leads to the equation:

$$(C_T^f u_f^2 + \alpha \Delta T g h)\frac{dh}{dt} = u_f^3 (1 - \frac{2\Phi}{u_f^3}) \qquad 13$$

where C_t^f is the efficiency with which water from the hypolimnion is entrained by falling plumes and mixed to the same state as the eplimnion; it is a calibration coefficient found by Fischer et al (1979) to equal 0.5. α is the thermal coefficient of expansion of water; ΔT is the change in temperature across the thermocline; h is the depth of the mixed zone; and Φ is equal to ε divided by the average density.

To calculate the energy balance for the typical wetland, with the initial thermal conditions taken as shown Figure 2, and other initial conditions assumed to be:

- an epilimnion 0.05 m deep;
- a temperature difference between the epilimnion and the top of the stratified water of 0.5°C; and
- a temperature gradient below the epilimnion of 4.4°C/m.

Given these initial conditions, times taken to destratify the hypothetical wetland to its full depth of 0.5 m under cooling rates of 50 and 100 Wm^{-2} are given in Table 6; calculated using a time step of 30 s. Under these conditions, it was found that the temperature drop across the thermocline quickly falls to zero, due to the heat loss from the wetland surface; when this occurs, destratification over the whole depth of the wetland proceeds almost instantaneously.

TABLE 6:
DESTRATIFICATION DUE TO PENETRATIVE CONVECTION
Initial mixed zone depth 0.05 m, initial thermocline temperature difference 0.5°C, temperature gradient of 4.4 °C/m in the unmixed zone,

Cooling Rate (W/m^2)	50	100
Fall Velocity (m/s)	0.0011	0.0014
Destratification Time (minutes)	35	18
Dispersion over destratified wetland (m^2/s)	0.0018	0.0021

Also presented in Table 6 are estimates of the dispersion coefficient due to penetrative convection. These are calculated by assuming that thermals due to penetrative convection will decay to form turbulent eddies on a length scale of approximately the wetland depth. Thus a dispersion coefficient describing the effects of these thermals on mixing over the wetland depth, after destratification has taken place is given by $K \propto u_f d$, where d is the wetland depth, say 0.5 m, u_f is the fall velocity of the thermal. The coefficient of proportionality should account for the efficiency given in Section 4.3, so the dispersion rate due to penetrative convection can be estimated as:

$$K \cong 0.3 u_f d \qquad 14$$

4.5. Mixing Due to Wind

A large and highly developed body of literature exists on momentum exchange between wind and water. The phenomena involved are highly complex, and a full presentation is beyond the scope of this report; rather scaling analyses will be used to provide a first estimate of the effects of momentum exchange from air to water that causes mixing in the water.

The depth averaged dispersion due to wind can be evaluated under neutrally buoyant conditions by scaling the dispersion coefficient with the shear friction velocity (u_*) and the characteristic length scale l over which the dispersion occurs. Assuming the efficiency of the process to be 30% as discussed earlier, results in:

$$K \approx 0.3u_*l \qquad 15$$

It is known that stratification lowers dispersion, so it is only necessary to evaluate K under neutral conditions to determine an upper bound of its value under stratified conditions.

Now, u_* is given by $\sqrt{\tau/\rho_o}$, where τ is the shear stress due to wind (Nm^{-2}), and ρ_o is the ambient water density (approximately 1000 kgm^{-3}). From Fischer et al (1979), this shear stress is:

$$\tau = \rho_a C_D U_a^{\,2} \qquad 16$$

where ρ_a is the density of air which is $1.2\ kgm^{-3}$ and C_D is the drag coefficient which is approximately 10^{-3}.

The length scale l in Equation 15 is the limiting dimension over which the turbulence occurs. Craig and Banner (1984) found this length scale to be approximately 10 times the amplitude of the surface waves induced by the wind; however, obviously the depth forms an upper limit on this value. The depth therefore provides a conservative estimate of the length scale; that is, an estimate that will overestimate the magnitude of the dispersion.

From these formulations for u_*, an estimate for the dispersion is given by:

$$K \sim 0.0003 U_a d \qquad 17$$

Table 7 presents surface water velocity and dispersion dependence on wind speed up to 10 m/s.

TABLE 7:
WIND INDUCED SHEAR VELOCITIES AND DISPERSION COEFFICIENTS

Assuming a 0.5 m depth of water

Wind Velocity (m/s)	0	1	3	10
Surface Velocity (m/s)	0	0.03	0.09	0.3
Dispersion (m^2/s)	0	0.00015	0.0005	0.002

4.6. Rain Effects

There is a substantial body of information about the kinetic energy of rainfall in meteorology literature. Rainfall energy fluxes over a range of rainfall rates from a North American study are presented in Table 8 (Maidment, 1993).

Studies examining rain falling on the ocean surface reveal that raindrops form vortex rings when they penetrate the water surface (Morton and Creswell, 1992). These vortex rings will decay to form turbulence. To parameterise this process, assume the turbulent velocity after the impact of the raindrop will be:

$$v_{turb} \approx \sqrt{\frac{2\,Energy\,/\,Area}{3\rho l}} \qquad 18$$

Taking $K \cong 0.3vl$, where K is the dispersion, l is the length scale and assuming 30% efficiency as in Section 4.3, gives:

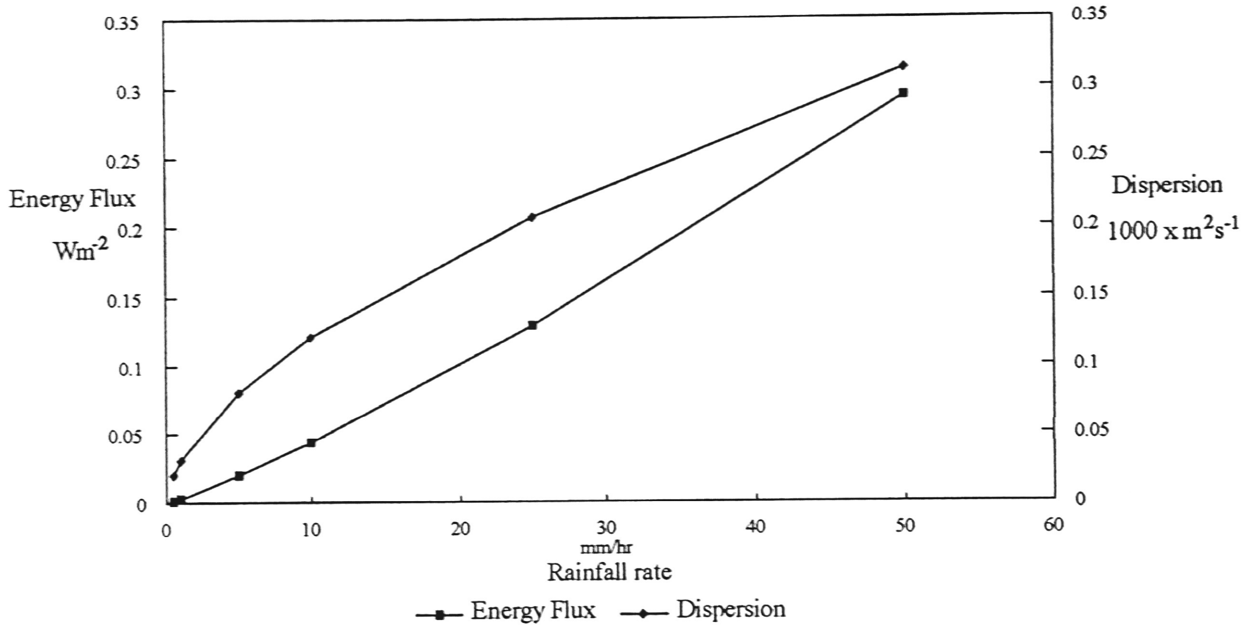
$$K_{rain} \cong 0.3\sqrt{\frac{2E_u l}{3\rho}} \qquad 19$$

Where E_u is the energy per unit area imparted by the rain on the water surface. From Morton and Creswell (1992), l is of the order of several centimetres (say 0.05 m), so dispersion can be calculated as a function of rainfall rate as presented in Table 8 and Figure 3.

TABLE 8:
RAINFALL ENERGY FLUX AND DISPERSION DEPENDENCE ON RAINFALL RATE

Rainfall Rate	mm/hr	0.5	1	5	10	25	50
Energy Flux	W/m^2	0.0013	0.0029	0.019	0.044	0.13	0.29
Upper bound of Dispersion	m^2/s	2×10^{-5}	3×10^{-5}	8×10^{-5}	1×10^{-4}	2×10^{-4}	3×10^{-4}

FIGURE 3:
ENERGY FLUX AND DISPERSION DEPENDENCE ON RAINFALL RATE



4.7. Effect of Mean Discharge through a Wetland

The magnitude of the dispersion coefficient due to the mean flow in a water body depends on whether the flow is laminar or turbulent on the scale of the mean flow. The mean flow Reynolds number is used to delineate whether the flow is turbulent or laminar (Vennard and Street, 1982). The mean Reynolds number is given by the equation:

$$R = \frac{ud}{\upsilon} \qquad (20)$$

From Table 2, for the hypothetical "typical" wetland; the mean velocity, u is 2×10^{-4} m/s, the depth d is 0.5 m and the kinematic viscosity of water, υ is approximately 1×10^{-6} m^2/s (Vennard and Street, 1982). This gives a Reynolds number of 100, which is well below the critical Reynolds number, below which open channel flows can be considered laminar.

In a laminar shear flow, the value of the dispersion coefficient is the rate of molecular diffusion (Fischer et al, 1979) so the values from Section 4.1 are appropriate.

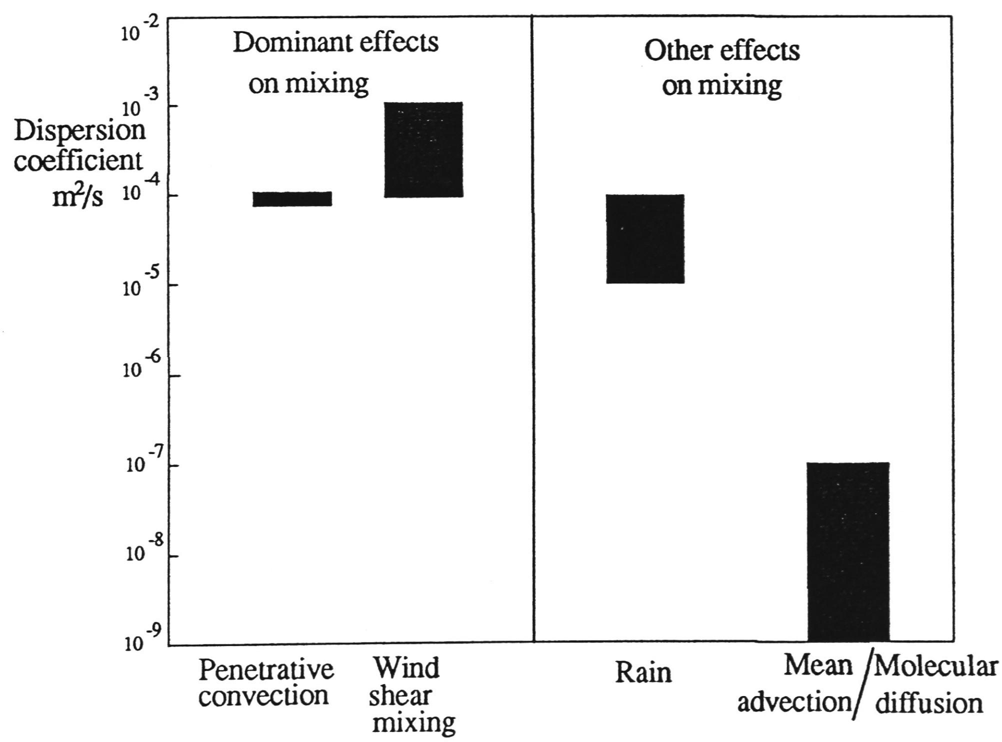
4.8. Summary of Dispersion Values

From the values of dispersion coefficients in Table 9 and Figure 4, when the major mixing processes are absent, dispersion must be greater than 10^{-7} m²/s, the value of molecular diffusion. Penetrative convection will increase dispersion to the order of 10^{-4} m²/s. Wind may increase dispersion to between 10^{-4} and 10^{-3} m²/s. While rain may give rise to dispersion values between 10^{-5} and 10^{-4} m²/s.

TABLE 9: VALUES OF DISPERSION COEFFICIENTS

Mechanism	Order of dispersion coefficients (m^2/s)	Order of turbulent velocity scales (m/s)
Molecular diffusion and Mean advection	10^{-9} for solutes to 10^{-7} for heat	-
Penetrative convection	10^{-4}	10^{-3}
Wind-shear effects	10^{-4} to 10^{-3}	10^{-1}
Rain effects	10^{-5} to 10^{-4}	-

FIGURE 4: RANGE OF DISPERSION VALUES OF MIXING MECHANISMS



5. DESTRATIFICATION DUE TO WIND MIXING

Having estimated the magnitudes of the expected stratifications and mixing processes in Sections 3 and 4 and found surface shear due to wind to be one of the dominant mixing processes, it is now useful to analyse the effect of this phenomenon on an existing stratification profile. This interplay between shear mixing and stratification is characterised by the Richardson number, which is a dimensionless parameter that compares the forces due to the density and velocity gradients within a water body being destratified by shear driven mixing. It is given by:

$$\Re_i = \frac{\frac{-g}{\rho}\frac{d\rho}{dz}}{(\frac{du}{dz})^2} \qquad 21$$

Consider the situation which may be expected on a still summer morning in a wetland that is initially unstratified at time $t = 0$. First the wetland becomes heated by solar radiation, with no mixing occurring, from an initial fully mixed condition, until a stratified condition applies at time t_1.

Stratification is caused by a temperature gradient in the water column, for this, the Richardson number can be expressed as:

$$\Re_i = \frac{\frac{-\alpha}{\rho}\frac{dT}{dz}g}{(\frac{du}{dz})^2} \qquad 22$$

where $\alpha \approx -0.24$ kgm^{-3}/°C. From Equation 4 in Section 3.2.1 for an initial condition at time $t = 0$ of a fully mixed water column, the temperature in the wetland after time t_1 over which no destratifying effects are present is given by:

$$T = \frac{-t_1}{\rho C_p}\tau_s \phi e^{-K_1 z}\{K_1[K_2 \tan^{-1}(K_3 z) - 1] - \frac{K_2 K_3}{1+(K_3 z)^2}\} \qquad 23$$

So that the temperature gradient at time t_1 is given by:

$$\frac{\partial T}{\partial z} = \frac{t_1}{\rho C_p}\tau_s \phi e^{-K_1 z}\{K_1^2[K_2 \tan^{-1}(K_3 z) - 1] + \frac{2K_2 K_3^2 z}{[1+(K_3 z)^2]^2}\} \qquad 24$$

Substituting typical values for the coefficients, K_1 = 0.2 m^{-1}, K_2 = 0.4 and K_3 = 4 m^{-1}, taking the surface transmissivity τ_s = 98% and evaluating the temperature gradient at the mid-depth of the water body, z = 0.25 m, results in Equation 25 as an expression for the temperature gradient over the whole wetland.

$$\left.\frac{\partial T}{\partial z}\right|_{t_1, z=0.25m} = \frac{0.77\phi t_1}{\rho C_p} \qquad 25$$

Now, assume that from time t_1 a wind with speed U_a begins to blow over the wetland. The velocity gradient will scale as the difference between the surface velocity at the surface and the velocity at the bed, divided by the depth h. The surface velocity is given by George (1974) as 3% of the wind speed, while the velocity at the bed must be zero. The velocity gradient is therefore given by:

$$\frac{dU}{dz} \approx \frac{\Delta U}{\Delta z} = \frac{0.03U_a}{h} \qquad 26$$

The approximation for the Richardson Number over the 0.5 m depth of the water body is then given by:

$$\Re_i \approx \Re_{io} = \frac{856\alpha\phi t_1 g h^2}{C_p \rho^2 U_a^{\ 2}} \qquad 27$$

From Turner (1973), the water column becomes unstable when the Richardson number exceeds a critical value given by:

$$\Re_i > \Re_i crit = 0.25 \qquad 28$$

This Richardson number criterion then leads to the expression in Equation 29 for the critical wind speed required to make the stratification in the water body unstable.

$$0.25 \sim \frac{856\alpha\phi t_1 g h^2}{C_p \rho^2 U_a^{\ 2}} \qquad 29$$

Substituting values for the constants and rearranging Equation 29 yields an expression for the critical wind speed as:

$$U_a crit \sim 0.0009\sqrt{\phi t_1} \qquad 30$$

where $U_a crit$ is in m/s, ϕ is in Wm^{-2} and t_1 is in seconds. Assuming the previously used thermal loading of 400 Wm^{-2}, minimum wind speeds required to destratify the wetlands for various times of stratification can be developed; these are presented in Table 10.

Below the critical wind speed, the stratification will remain stable, so that a Richardson number correction will be needed for dispersion, and stratification will significantly affect the rates of mixing in the wetland.

TABLE 10: WIND SPEEDS REQUIRED FOR DESTRATIFICATION

for a Thermal Loading of 400 Wm^{-2} and wetland depth of 0.5 m

Stratification time (hours)	1	2	5	10
Minimum Wind Speed for Destratification (m/s)	1.0	1.5	2.4	3.4

6. INFLOW EFFECTS

The momentum of the water entering a wetland will be the dominant mixing mechanism over some area of the wetland around the inlet. The length scales defining this area of the wetland are determined by the density and momentum differences between the inflowing water and the water already in the wetland.

6.1. Inflow Buoyancy

Table 11 presents the most extreme temperature induced density differences recorded for West Byron, for January to July 1992 as reported by Bavor et al (1992).

TABLE 11:
PROPERTIES OF INFLUENT RELATIVE TO AMBIENT WATER AT WEST BYRON WETLANDS

	ΔT (°C)	$\Delta \rho$ (kgm^{-3})
Maximum	3.0	1.2
Minimum	–2.7	–1.1

For the case where the inflow is warmer than the fluid within the wetland, the inflow will ride over the top of the ambient water. This will affect the hydraulic behaviour of the wetland in the following ways.

- Short circuiting may occur if the wind is blowing from inlet to outlet, as wind driven surface velocities will operate on the overlying, inflowing water, driving it towards the outlet, while the ambient fluid would be largely unaffected.

- Stratification may be enhanced by the warmer inflowing water overlying the cooler ambient water which reduces vertical and horizontal mixing.

Alternatively, where the inflow is cooler than the fluid within the wetland, the inflow will form an underflow along the bed of the wetland. It is also possible for an interflow, or intrusive flow, to form if the fluid within the wetland is stably stratified and the temperature of the inflowing water is equal to the temperature of the ambient water at a particular depth.

6.2. Turbulent Buoyant Jet Analysis

Assuming that water is discharged into the wetland from a single, horizontal, submerged pipe, which is remote from the wall so there are no wall effects on the discharge. This will ensure the effects of the inflow are projected as far into the wetland as possible, providing an upper bound estimate of the spatial extent of the wetland affected by the inflow.

For the case of the previously assumed inflow Q = 0.003 m^3/s, assume that the water discharges through a 50 mm diameter pipe with an average influent velocity of 1.5 m/s and the influent discharge Reynolds number is 8 000. When the Reynolds number of a jet exceeds 4 000 the jet is turbulent (Fischer et al, 1979); so the following analysis, which assumes turbulent conditions is valid.

Fischer et al (1979) defines length scales for a circular buoyant jet as:

$$l_m = \frac{\left[\frac{\pi D^2}{4} v^2\right]^{\frac{3}{4}}}{\sqrt{\frac{\pi D^2}{4}\frac{\Delta\rho}{\rho_0} g v}} \qquad 31$$

$$l_{mu} = \frac{\left[\frac{\pi D^2}{4} v^2\right]^{\frac{1}{2}}}{U} \qquad 32$$

where v is the velocity of the incoming fluid, U is the velocity of cross flows, D is the inlet diameter, $\Delta\rho$ is the density difference between inflowing and ambient waters, ρ_o is the ambient density, l_m is the distance over which momentum dominates buoyancy and l_{mu} is the distance over which momentum dominates cross flows. Tables 12 and 13 present values for these length scales over a range of density difference conditions, assuming the cross flow is wind generated and is determined by $0.03U_a$.

TABLE 12:
MOMENTUM – BUOYANCY LENGTH SCALES

$\Delta\rho$ (kgm^{-3})	l_m (m)
1.2	16
0.5	80
0	∞
–0.5	80
–1.1	17

TABLE 13:
MOMENTUM – CROSS FLOW LENGTH SCALES

Wind Speed (m/s)	l_{mu} (m)
1	2
3	0.7
10	0.2

Clearly the buoyancy effects in this case are quite small; l_m is large, indicating that the inflow behaves largely as a horizontal, momentum dominated jet. Only slight winds are required for ambient water movement to easily dominate over the most severe inflow momentum effects; hence the inflow effects will have only a minor influence on the overall wetland behaviour.

Jet width and thickness for a simple turbulent axi-symmetric jet discharging into a like fluid as presented in Fischer et al (1979) are:

$$b_u = 0.107x \qquad (33)$$

where x is the distance downstream of the inlet point. Thus the distance to the point where the jet would encompass the whole depth of a wetland 0.5 m deep is 4.7 m. It is therefore apparent that wind driven effects will dominate over inflow effects well before inflow effects have an influence on mixing in the wetland.

Without a more detailed analysis, the following conclusions can be drawn.

- Inflow discharges into the wetland will be momentum dominated, and are likely to act according to bounded jet theory for the most part; however, the jet effects do not carry far into the wetland, being easily dominated by wind driven motions.

- Inflow effects will be very dependent on the configuration adopted and are thus highly site specific. Changes in pipe diameter, the use of inclined inflows, multiple diffusers, plunging weirs and vegetation will all significantly affect the degree of mixing immediately around the inflow and the extent of this inflow mixing zone.

- The choice of a particular inflow arrangement is highly dependent on what is considered desirable in wetland design; if stratification and turbulence are considered undesirable, then the use of multiple diffusers discharging vertically would be appropriate.

7. WITHDRAWAL EFFECTS

Withdrawal refers to outflow effects observed in stratified water bodies, whereby vertical buoyant forces are sufficiently strong that the outflowing water comes from a thin horizontal layer, at the level of the discharge point (Fischer et al 1979).

A v-notch weir outlet is an example of axi-symmetric withdrawal, that is, withdrawal from a three-dimensional water body by a point sink. Ivey and Blake (1985) provide a comprehensive treatment of this case. They present formulae for the half thickness of the layer from which water is withdrawn in the water body, for flow regimes determined by values of the following dimensionless parameters:

- a transition parameter: $$S = \left(\frac{Q^2 N}{\nu}\right)^{\frac{1}{15}} \qquad 34$$
- the Grashof number: $$G_r = \frac{N^2 L^4}{\nu^2} \qquad 35$$
- the Schmidt Number: $$S_c = \frac{\nu}{D} \qquad 36$$

where N is the buoyancy frequency, in radians per second, given by Equation 37, D is the diffusivity of the stratifying constituent and L is the limiting horizontal length scale, which in this case is the width of the wetland.

$$N = \sqrt{\frac{-g}{\rho_o}\frac{d\rho}{dz}} \qquad 37$$

For the density regime presented previously in Table 3, it can be shown that $N = 0.1$ radians per second.

The significance of these non-dimensional parameters is:

- S determines whether the flow is dominated by viscosity or inertia
- G_r compares the importance of buoyancy influences to viscosity and
- S_c compares viscosity to diffusion.

Table 14 presents values of S and G_r over the range of values expected in wetlands. Note that S_c will be approximately constant. In this context, it is more appropriate to use the dispersion values derived in Section 4.7, rather than the molecular diffusion, as turbulent dispersion will dominate over molecular diffusion in this case. If from Section 4.8, D is 10^{-4} m²/s, and ν is 10^{-6} m²/s, then S_c will be 0.01.

TABLE 14:
PARAMETERS IMPORTANT TO WITHDRAWAL OVER THE RANGE OF EXPECTED CONDITIONS
with N = 0.1 Hz, $\nu = 10^{-6}$ m²/s, D = 10^{-4} m²/s

	Q (m³/s)	L (m)	S	G_r	S_{lim}	δ
Smallest wetland	5×10^{-5}	1	3.6	1×10^{10}	6	0.05
Typical wetland	3×10^{-3}	40	6.3	3×10^{16}	14	0.16
Largest wetland	10	500	18	6×10^{20}	24	0.38

Ivey and Blake (1985) conducted experiments showing that at a critical value of S, $S_{crit} \approx 3$, above which the flow regime is inertial-buoyant, and below which the withdrawal flow regime is viscous-buoyant. As shown in Table 14, over the range of conditions expected in wetlands, flows are inertial-buoyant. Ivey and Blake (1985) distinguish two subclasses, depending on the magnitude of S relative to G_r and S_c. The separation between these regimes is given by:

$$S_{\lim} = (G_r S_c^{-2})^{\frac{1}{18}} \qquad 38$$

Values of $S_{\lim}$ are also given in Table 14; from these values it can be seen that withdrawal from wetlands always falls into the regime $S < S_{\lim}$ and $S > S_{crit}$, thus they can be classified as inertial-buoyant. For this situation, Ivey and Blake (1985) give the value of the half thickness of the withdrawal layer δ as:

$$\delta = c_2 (\nu D)^{\frac{1}{6}} (\frac{L}{N})^{\frac{1}{3}} \qquad 39$$

where c_2 has a value of 2.9. Values of δ are given in Table 14 from which, the typical half thickness of the withdrawal layer is 0.16 m. This thickness is comparable to the total depth of the wetland, so that withdrawn water is not limited to a small layer in the wetland. Consequently withdrawal effects are not expected to be significant.

8. CONCLUSION

External influences on mixing processes in wetlands have been examined through scaling arguments. It has been found that the most significant external influences on mixing are:

- direct wind; and
- penetrative convection

Rainfall may play a significant role when it occurs. The flow is laminar in the mean, so the dispersion due to the mean flow is equal to the molecular diffusion.

The phenomena that dominate mixing are expected to give rise to dispersion coefficients ranging from 10^{-5} to 10^{-2} m^2/s, and are capable of destratifying wetlands in which even the most severe stratification has been established; however, these phenomena will only be present intermittently.

The degree of stratification that develops will greatly influence the effectiveness of both the stronger and lesser mixing influences. Strong stratifications of up to 2.5°C over a typical wetland depth of 0.5 m are expected in summer; markedly different summer and winter conditions are therefore expected. Stratification due to solar radiation can be established in a few hours, but breaks down rapidly due to penetrative convection and wind speeds in excess of 3 m/s. Inflow and outflow mixing processes are likely to have only limited effect on the overall transport processes.

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