THE APPLICATION OF A DYNAMICS SIMULATION MODEL TO LAKES AND RESERVOIRS

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ABSTRACT

A process oriented numerical model is applied to two diverse enclosed water bodies for the simulation of their lake wide average temperature and salinity distributions in the vertical direction over long term periods of several years. The principal dynamical processes accounted for in the model are the surface mixed layer formation including the stirring action of the surface wind stress, the velocity shear at the base of the mixed layer due to internal waves and the buoyancy flux due to surface heat exchanges, billowing at the base of the mixed layer, inflows of various densities and surface and subsurface outflow. Deep hypolimnetic mixing is accounted for by vertical diffusion following the theory of Obukov.

The assumption of one-dimensionality is discussed in terms of three non-dimensional numbers, the Wedderburn number, the ratio of the Rossby radius to the basin breadth and the internal Froude numbers associated with the inflows and outflows. While this analysis shows that one body, a reservoir, is dominated by surface mixing and bottom inflow and the other, an intermediate scale deep lake, is dominated by shear induced mixing and billowing and deep hypolimnetic mixing, both lakes behave on the whole essentially as one-dimensional in the vertical. Model hindcasts replicate all the main features of the field data of both cases and on a quantitative basis the volume weighted or RMS temperature error is 1°C and salinity from 150 to 4 ppm depending on the water body.

The application of the model to the problem of interpolation between field surveys is illustrated by the example of a resistent winter temperature inversion supported by a weak salinity gradient. The times of formation and erosion of this feature predicted by the model are supported
by the observations of Wiegand and Carmack (1981). The independent calibration of model coefficients and the critical comparison of model output to field observations has indicated that the principal shortcoming of the model is its ability to account for mixing in the hypolimnion on a seasonal time scale or longer.
1. INTRODUCTION

Interest in the simulation of the behaviour of lakes and reservoirs has arisen from a need to optimally manage the resource. Simulation models are being applied to such problems as determining long term water quality trends and the response of lake ecosystems to various management strategies. These goals are distinct from those of specific models of such dynamical phenomena as lake circulation and storm surges, although such phenomena may well be incorporated into a long term model.

Many long time scale models have been built with the aim of simulating water quality (see Orlob, 1977 for a review); frequently these have relied on the calibration of empirical eddy diffusivities to predict the thermal structure, assuming that modelling of water quality parameters will directly follow. The difficulty with this approach is that different water quality parameters may depend on different mixing processes for their distribution (Fischer et al, 1979); without specific inclusion of these processes there can be no guarantee that the simulations are realistic. Further, such calibrated models yield little or no insight into the dynamics of the lake and the interactions between the various processes, and the transfer from one lake to another involves a recalibration of the various model coefficients.

On the other hand, the inclusion of the parameterisations of the known processes governing the lake dynamics results in increasingly complex models which, notwithstanding problems of numerical accuracy, would realistically be able to simulate only short periods of real time. However, in such process oriented models, parameter values inherent in the descriptions should remain constant from lake to lake and be independent of the
range of input data, and in many cases their values are obtainable independently of the modelling exercise.

Clearly such process oriented models have a number of advantages beyond those above; the application to widely varying input data will be more reliable, providing a means of testing the effects of changing external inputs; such models are capable of providing a visualization of specific events, an identification of the individual processes involved, and the reason for their occurrence; inconsistencies in limnological data time series may be detected, and missing data reconstructed to provide a full description of the dynamics of the lake.

The complexity of these models remains as the major difficulty. However, the strong stratification normally found in small to medium sized lakes and reservoirs may be used to advantage. Vertical motions are inhibited and turbulence is reduced to intermittent bursts. As a result, longitudinal and transverse variations play a secondary role with variations in the vertical being the most important contributions to the first order balances of mass, momentum, and energy. A one-dimensional description is consequently applicable, with the ensuing advantage that the required temporal and spatial resolutions are attainable.

In this paper we describe the application of such a one-dimensional process oriented model to two diverse water bodies. The model, DYRESM, was first described by Imberger et al (1978), and in its current form by Imberger and Patterson (1981). The model makes use of a movable representation of a lake, the temporal and spatial resolutions being varied by the algorithm as the requirements of accuracy and efficiency dictate for each process modelled.
In order to fully test its capabilities, the model has been applied to a small, shallow reservoir, Wellington Reservoir in Western Australia, and to an intermediate scale, deep montane lake, Kootenay Lake in British Columbia, without recalibration of the model constants and only minor changes to the physical process description. These two water bodies were chosen as subjects for the test as both had been the subject of intensive field studies, and good data sets exist for input to and validation of the model. Further, as the horizontal scale of the water body increases it may be expected that the assumption of one-dimensionality will become less valid; in this sense Wellington Reservoir and Kootenay Lake represent a wide range of behaviour, and so provide a good test of the transferability and predictive capability of the model.

In the following, we briefly describe the field programs undertaken on the reservoir and the lake and discuss in some detail the principal dynamical processes which may be inferred from the field observations. This insight into the dynamics prior to modelling leads to an assessment of the validity of the one-dimensional assumption. The model DYRESM is briefly described, and results of lengthy simulations on each water body are given and critically compared with the field data and some model deficiencies noted. The conclusion drawn is that the model performed satisfactorily on both lakes, and has the additional benefit of giving insight into details of the various processes.
2. THE FIELD PROGRAMMES

Wellington Reservoir

Wellington Reservoir is a 185 x 10^6 m^3 capacity storage located approximately 150 km south of Perth in Western Australia. The reservoir is approximately 30 m deep at the dam wall and extends some 24 km along the Collie River valley; at crest level the surface area is approximately 160 ha (Figure 1). The climate of the region is Mediterranean in nature and rainfall and streamflow are strongly seasonal. Draw from the reservoir is primarily for irrigation use during the summer period October to May, although a small component for domestic use is maintained throughout the year. During winter and early summer, water is drawn from a deep offtake as part of a management programme.

Disturbance of the catchment by clearing for agricultural use has led to significantly increased salinities in the inflowing streams, and the resulting salinity build up in the reservoir led to the initiation, in 1974, of an intensive study of all aspects of the reservoir behaviour. Development of the simulation model DYRESM commenced during this study, as did an intensive field monitoring program, which continued until early 1978. The full results of the field study have been reported elsewhere (Imberger et al, 1978, Hebbert et al 1979, Imberger and Hebbert, 1980) and only brief details will be given here.

The field program consisted of obtaining temperature, conductivity, and (sometimes) dissolved oxygen profiles at each of the stations indicated on Figure 1 at approximately one month intervals, commencing in late 1974 and continuing to December 1977. The actual visitation frequency to the site or to individual stations depended on the observed phenomena,
hence there are occasional gaps of more than one month in the field data and conversely there are several periods of intense data.

Whole lake averaged profiles of temperature and salinity, obtained from a locally developed empirical formula relating conductivity and salinity (Imberger and Hebbert, 1980), were constructed for each day on which field data were taken. These mean profiles span the simulation period of 964 days, May 11, 1975 (day 75133) to December 31, 1977 (day 77365).

The inputs forcing the reservoir were also measured during this period; the variations of the major river inflow volume, temperature and salinity, together with the solar radiation intensity and daily average wind speeds (> 3 ms⁻¹) are shown in Figure 2, which demonstrates the seasonal nature of the forcing. The major inflows occur during the winter months, July to September, and are cold, and in the early part, highly saline. This is also a period of minimum surface heating and high wind activity. The effect of this forcing on the reservoir is shown in Figures 3a and 3b, which give the isotherm and isohaline depths as a function of time, constructed from the mean profiles.

The cycle of events is clear. At the beginning of the period, the reservoir is temperature stratified, with only a weak salinity gradient. Following the winter overturn, the cold, salty inflows lodge in the base of the reservoir, which then restratifies during the following summer. Clearly the bottom temperature is determined principally by the temperature of the coldest inflow, whereas the surface temperature is determined by the meteorological inputs. The degree of stratification is somewhat smaller in the 1977/78 summer than in the preceding summer; this was due
to a changed operational policy, whereby large quantities of water were drawn from a deep offtake in an attempt to minimize the salinity build-up in the reservoir.

This heavy withdrawal of bottom water during a period of high stratification had the effect of dismantling the density structure, with the result that the winter overturn was advanced by several months in the following year.

In interpreting Figures 3 it needs to be noted that the dashed lines on Figure 3a have been constructed from spot profiles taken in the main basin of the dam. The error associated with this procedure is thought to be small, as the main basin is representative of the whole reservoir.

**Kootenay Lake**

Kootenay Lake is a long (107 km), narrow (4 km), and deep (mean depth 100 m) fjordic lake located in southeastern British Columbia. The maximum depth of the lake is 0(150 m) and its surface area 0(3.6 x 10^4 ha). The lake consists of a north and south arm with a sill outlet located at the junction of the arms (Figure 4). Although the lake is situated at 50°N latitude, the surface remains ice free.

An extensive limnological study was initiated in 1976 to examine the effects of upstream impoundments on the environmental conditions in the lake. During the period June 1976 to November 1977, fifteen whole lake surveys were conducted at approximately monthly intervals. Among a number of limnological properties temperature and conductivity were measured at the station locations shown in Figure 4. At each station continuous profiles of temperature and conductivity were measured and whole lake
averages of temperature and conductivity at 15 standard depths, computed by a volumetric weighting procedure, formed the basic data set simulated in this study. Salinity was evaluated from the measured conductivities through an empirical relation (Hamblin and Carmack, 1980).

Full details of the survey program and a full description of the data are given in Daley et al, 1990, and associated reports. The simulation period covered the extent of the field program from June 16, 1976 (day 76168) to November 16, 1977 (day 77320), a total of 519 days.

The nature of the forcing is shown in Figure 5, which gives the variation of North Arm inflow volume, and temperature and salinity of the North and South Arm inflows, together with the solar radiation intensity and daily average wind speeds (> 3 m s⁻¹) for the simulation period.

The seasonal trend is not as clear as in the Wellington Reservoir, however, the May freshet period and the autumn and winter flows are relatively high due to upstream power generation demands. The lack of seasonality is of course due to the upstream regulation of both rivers. Compared to long term averages 1976 was a wet year with a higher than normal inflow during the May freshet period while 1977 was a drier than normal year. Two-thirds of the ungauged river inflow as determined from mass balance, (Daley et al, 1980) was assigned to the South Arm with the remainder going to the North Arm and in both cases these inputs were assigned the conductivity and temperature of the inflowing rivers. Also worthy of note is the nature of the inflow salinity data; this has been interpolated from spot monthly values and may be seriously in error, particularly during times of high flows.
The effects of the forcing may be seen in Figures 6a and 6b which show the history of the isotherm and isohaline depths, calculated from the whole lake profiles, over the simulation period. The period of summer stratification extends from June to October, with weak temperature gradients in both epilimnion and hypolimnion. By the end of October rapid cooling begins and the weak salinity gradient becomes the dominating stratifying agent as the temperature of maximum density is approached. Consequently, a temperature inversion has appeared during January, shown by the 4°C, and 4.25°C isotherms on Figure 6a. The inversion is weak, at most 0.5°C, and there is no distinct mixed layer. The inversion has disappeared by the end of March, when stratification recommences. The second summer period is similar to the first, with a somewhat more distinct thermocline.

In interpreting Figures 6a and 6b it should be noted that the isotherm and isohaline positions are interpolated from the approximately monthly mean profiles. Thus events which occur over time scales less than this will be smoothed temporally in Figure 6. This is particularly true of the November/December 1976 mixing events, when a gap of some two months occurs between the full field surveys on November 2 and January 25 the following year. In the absence of other data, the interpolation smooths the data which appears as mixing over a lengthy period, as shown in Figure 6. Spot surveys were done on December 14 and January 1; these data indicate isotherms histories which show a much sharper rise to the surface at this time (Daley et al, 1980).

The same comment applies to the temperature inversion in January-March, 1977. Two regular surveys were carried out during the period of existence of the inversion, and coupled with the fact that the overall temperature
difference between surface and bottom is less than 0.5°C, this means that the depiction of the inversion is very much an interpretation of the interpolation procedure. Its presence however is confirmed by the fixed temperature profilers, as discussed later.

One further comment is also applicable to Figures 6a and 6b. Daley et al (1980) reported considerable internal wave activity; the effect of averaging the profiles taken over the length of the lake over a period of up to three days will be therefore to spread any sharp gradients of temperature and salinity vertically. Thus the lack of sharpness of the interface evident in Figures 6a and 6b may in part be due to this averaging process. An estimate of this effect will be made in the following section.
3. DYNAMICAL CONSIDERATIONS

As a prelude to the simulation of the two lakes, it is instructive to examine the field data with a view to establishing the role of the various mechanisms which dominate the behaviour and determine the resulting mean profiles. This procedure also highlights the major differences between the lakes in terms of their dynamics and provides criteria for the use of a one-dimensional model. The mechanisms considered here are the mixed layer dynamics, hypolimnetic mixing, inflow dynamics, and outflow dynamics.

**Mixed layer dynamics**

The depth of the surface mixed layer is determined principally by the stirring action of the surface wind stress, the velocity shear in and at the base of the mixed layer, and the buoyancy flux resulting from surface heat exchanges. The relative importance of the stirring and shear production mechanisms in a lake of two layer character may be gauged from the value of the Wedderburn number (Imberger and Hamblin, 1982)

$$W = \frac{\alpha' h h}{u^* L},$$

first introduced by Thompson and Imberger (1980), where $h$ is the mixed layer depth, $L$ the reservoir length, $g'$ reduced gravity across the interface, and $u^*$ the wind shear velocity. For $W \sim O(1)$, the interface surfaces at the upwind end; for $W >> O(1)$, i.e. low winds and strong stratification, the interface motions are small and mixed layer deepening is dominated by stirring (Spigel and Imberger, 1980). As $W$ decreases, the interface motions increase and shear production become increasingly important. Thompson and Imberger (1980) showed by numerical experiment that for $W < O(3)$, deepening is via a downwind movement of an upwelling
front. For $W < O(1)$, the deepening proceeds rapidly and the lake becomes homogeneous.

The shear production mechanism generated by the seiching of the interface acts until such time as the influence of the boundaries is felt in the form of a reflected seiche. Spigel and Imberger (1980) showed that this occurred in a time $T_i/4$, where $T_i$ is the seiche period, assuming a constant wind.

The value of $W$ then provides an indication of the type of mixed layer deepening expected from the effects of a surface wind stress. For $W >> O(1)$, stirring is expected to be the dominant mode of deepening, and the resultant interface will be sharp and the dynamics essentially one-dimensional. For $O(10) > W > O(3)$, shear production and billowing will become increasingly important, subject to production being cut off at times greater than $T_i/4$ or the duration of the wind event, whichever is the shorter. For $W < O(3)$, deepening by the movement of an upwelling front is likely.

The lake wide averaged temperature and salinity profiles generated from the field experiments provide a means of computing $W$ values and seiche periods for those days on which field data are available. As the parameterisation is based on a two-layer structure, it is necessary to construct such a structure which is equivalent in some sense to the lake wide averaged profile for each of the available profiles. Since the seiche period should be identified with the first mode internal wave period, a consistent means of constructing a two-layer structure is to locate the interface at the level of maximum density gradient in the lake wide average profile, and determine the density jump by matching the two-layer seiche
period (Turner, 1973) with the first mode internal wave period, both based on a rectangular basin. The results of this procedure are shown by the symbols in Figures 7a and 8a as computed W values for the days on which field data are available. The solid lines in Figures 8 and 9 are the result of the simulation, referred to later in this paper.

For the smaller lake, Wellington Reservoir (Figure 7a), it is clear that \( W > 0(10) \) for most of the simulation period, and is frequently \( 0(10^2) \). On two occasions, day 76198 (July 16) and day 76221 (August 8, 1976), \( W \) falls below 1 (0.77 and 0.62 respectively). In the first case, the effective \( g' \sim 0.2 \times 10^{-1} \text{ms}^{-2} \) at an interface depth of only 1.8 m; the average wind speed is 1.9 m/s yieldu* \( \sim 2.4 \times 10^{-3} \text{ms}^{-1} \). Although the wind speed is relatively low, the interface is sufficiently close to the surface to be significantly affected, and significant deepening will occur via the upwelling mechanism; indeed on the following day, the equivalent mixed layer depth has become 20.4 m. In the second case (August 8, 1976) the effective \( g' \sim 0.67 \times 10^{-2} \text{ms}^{-1} \) at an interface depth of 7.2 m; the wind speed of 4.8 m/s yields \( u* \sim 6.1 \times 10^{-3} \text{ms}^{-1} \). In this case the combination of a relatively high wind speed and weak interface yield a rapid deepening. In both cases, however, reference to Figure 3 shows that at these times the reservoir is very weakly stratified and the influence of these rapid deepening events on a mean profile which is nearly uniform is minimal.

The same is true for those occasions when \( W < 0(10) \); the low values of \( W \) are generally due to the weak stratification and the mean profile in these periods is only slightly modified. Thus, in general, although deepening by shear production and an upwelling front may very occasionally be in evidence, the effect on a lake wide average profile which has very
little structure will be minimal. In this sense, shear production, billowing, and upwelling will be significant to the mean structure only on those occasions when there is a well defined, shallow mixed layer and high winds. In the Wellington Reservoir, where the average wind speeds rarely rise above 5 ms$^{-1}$, this situation is rare. The normal circumstance is for $W > 0(10^2)$ and the principal effect of the wind on the mixed layer dynamics is via the stirring mechanism. This process, together with deepening by buoyancy flux, will result in a sharp, virtually horizontal interface, with little evidence of shear instability.

On the other hand, the computed values of $W$ for Kootenay Lake (Figure 8a) yield a different result. In this case $W$ is generally $< 0(10)$ with a few excursions into the region $W < 0(1)$. In general, the low $W$ values are the result of consistently high wind speeds (Figure 5) as well as the greater length of the lake. These values of $W < 0(10)$ indicate that shear production and billowing will play a major role in the development of the mixed layer, and will be associated with significant interface motions.

The $W < 0(1)$ condition arises on day 76168 (June 16, 1976), day 77089 (March 30, 1977) and day 77147 (May 27s, 1977), with $W$ values of .4, .9 and .29 respectively. The first and third of these events occur as the result of shallow interfaces ($h \sim 0(5m)$); in both cases $g'$ is $0(10^{-1})$. However $h/L \sim 10^{-5}$ and even moderate wind speeds are sufficient to cause a rapid deepening. The second occasion (March 30, 1977) occurs when $g' \sim 10^{-3}$; the interface depth is $0(75m)$ and is rapidly deepened to the bottom by the action of the wind.

Both these classes of low $W$ (strong but shallow and weak but deep
interfaces) also occur in the remainder of the $W$ values. There is, however, no evidence that the upwelling front predicted by Thompson and Imberger (1980) is present. This is because the seiche period $T_i \sim O(20 \text{ days})$ (on March 30, 1977 for example) yields $T_i/4$ much greater than the duration of a wind event, $O(6 \text{ hours})$. As a result, the interface does not surface at the upwind end before the wind event ceases or changes direction. Shear production will be cut off at the end of the event, although large interface motions may still be present (Daley et al, 1980).

The field data then indicate that in Kootenay Lake the mixed layer deepening will be largely controlled by shear production of turbulent kinetic energy at the interface, with smearing of the interface by Kelvin-Helmholtz billowing. Large scale internal wave motions may be expected, but the progressive upwelling front for $W < O(3)$ is unlikely to appear.

An estimate of the smearing of the lake wide averaged profile taken over several days may be obtained by considering a balance between the applied wind stress the pressure force resulting from the tilted interface. Thus, using (1), the peak interface excursion $d$ is

$$d \sim \frac{h}{W^2}$$

In Kootenay Lake $d \sim O(10 \text{m})$ for periods of strong stratification ($g' \sim O(10^{-1})$) and $d \sim O(50 \text{m})$ if the stratification is weak ($g' \sim O(10^{-3})$). In the Wellington Reservoir, $d \sim O(1 \text{ m})$ for the whole simulation period. The process of length averaging lake wide data then has little effect in Wellington Reservoir, but will smear the representation of the interface in Kootenay Lake over depths $O(10 \text{ m})$. 
A further effect which may influence the mixed layer depth in both lakes is that of Coriolis forces. Deformation of the interface will occur if the Rossby radius of deformation, $R$, is less than the width scale of the lake, $B$,

$$R = \frac{(g' h)^{1/2}}{f} < B$$

where $f$ is the Coriolis parameter. In figures 7b and 8b the symbols show the values of $R$ computed for each of the field data days. The width of the main basin of Wellington Reservoir is 0(2 km); reference to Figure 7b shows that $R$ rarely falls below this value. Comparison of Figure 7b and Figure 3 shows that on those occasions that $R < 0(2)$ (March 1975, June/July 1976, and February 1977), the stratification is extremely weak and hence $g'$ is small. Since there is effectively no interface at these times, the effect of Coriolis forces will not be evident. The width of Kootenay Lake at interface depth is 0(2 km) also; Figure 8b shows that $R$ is less than this value only during the winter period December to March, at which time the stratification is extremely weak. Again, it is unlikely that the influence of Coriolis forces on the interface will be evident.

In summary, it is clear that the mixed layer dynamics in the two lakes are dominated by different processes. However, in both cases, one-dimensional parameterisations for the processes are appropriate, and the simulation of the mixed layer behaviour will provide an excellent comparative test of the parameterisations.

**Hypolimnetic mixing**

The vertical transports in the hypolimnion are evidently due to a combination of mechanisms, including mixing on the boundaries, double
diffusive activity, intrusions originating from inflow, outflow or unequal heat capture, and local internal overturning events (Imberger and Hamblin, 1982). No clear parameterisation covering all these mechanisms appears to be available; the parameterisation suggested by Imberger et al (1978) however provides a cover for the estimated values of eddy diffusivities, and may be identified with other parameterisations (c.f. Garrett, 1979. Weinstock, 1978).

The Imberger et al (1978) model is based on the assumption that all processes in the lake are in equilibrium with the external forcing, and results in an eddy diffusivity coefficient formulation

$$K_z = \frac{0.048H^2}{T_M^M}$$

(4)

where $H$ is the lake depth, $T_M$ the potential energy stored in the lake stratification divided by the power input from the wind and the river inflow, and $S$ the stratification stability, $S = MN^2/g'$ where $N$ is the buoyancy frequency.

There are two parameters implicit in this formulation; firstly the parameter 0.048 which may be expressed as an efficiency factor for the conversion of the energy input into active mixing events, and secondly, the limiting value of $K_z$ approached as the stratification decreases. The efficiency factor associated with the parameter 0.048 may be given values ranging between 0.1% and 5.0%, depending on the shape of the lake basin, comparing favourably with other estimates (Imberger and Hamblin, 1982). The limiting value of $K_z$ is chosen to encompass the range of values discovered experimentally as well as to provide the correct model behaviour; as such it is the only parameter capable of calibration in the
present model. The value chosen as the limiting value of $K_z$ is $10^{-5} \text{m}^2 \text{s}^{-1}$.

In the Wellington Reservoir the bottom water replenishment is via the underflowing of cold, salty waters during winter, with the result that a salty wedge lodges in the base of the reservoir. The transport of this salt concentration to the upper regions occurs over a scale of months, and provides a measure of the mixing processes in the hypolimnion. The reservoir has a hypolimnion depth $O(20 \text{ m})$, with relatively gently sloping sides. On the other hand, Kootenay Lake has sharply sloping sides and a hypolimnion depth $O(100 \text{ m})$; the greater depth places more reliance on the correct parameterisation for vertical transport than in the case of Wellington Reservoir, where surface events regularly completely mix the reservoir. In Kootenay Lake however, bottom replenishment only occasionally occurs via vertical overturn. Thus in the Wellington Reservoir, hypolimnetic mixing provides only a part of the vertical transport, in contrast to the case in Kootenay Lake, where it provides the bulk of vertical transport. The great differences in hypolimnion depth, together with the different basin shapes, will therefore provide different bases for a test of the parameterisation.

Inflows

A river entering a reservoir or lake will either overflow the storage if it is less dense than the stored water or underflow to the bottom or some intermediate depth if it is denser than all or some of the stored water. The dynamics of the plunge line and entrainment of an underflowing stream have been analysed by Hebbert et al (1979) and Fischer et al (1979); the intrusion of the inflow at the level of neutral buoyancy has been parameterised by Imberger et al (1976). The process of cabling has been described by Carmack (1979), and Daley et al (1980).
In the Wellington Reservoir the cold salty winter inflows (Figure 2) underflow the stored water and lodge in the base of the reservoir (Figure 3b), providing the mechanism for bottom replenishment (Imberger and Hebbert, 1980). The winter flows into Kootenay Lake however usually enter at some mid-depth and only occasionally plunge. Cabbeling sometimes occurs, and provides the means of bottom replenishment for the lake (Daley et al., 1980). The internal Froude number, $F_i$, where

$$ F_i = \frac{U}{\sqrt{g'rH}} $$

(5)

where $U$ is the inflow velocity, $g'$ reduced gravity between inflow and surface water, and $H$ the total depth, provides a measure of the influence of the inflows; for both lakes this number is small. The values of $F_i$ computed from the field data are shown as symbols on Figures 7c and 8c. In their intensive study of the inflow to the Wellington Reservoir during the 1976 winter, Hebbert et al (1979) showed that $F_i$ (based on the hydraulic depth $H/2$) was typically $O(0.2)$, which corresponds to the $O(0.05)$ values for the same period shown in Figure 7c. Figure 8c shows that the $F_i$ values computed from the Kootenay Lake field data are $< 1$ for the whole period. The interpretation of these $F_i$ values is that, for both lakes the inflows are not sufficiently large to upset the one-dimensionality of the mean structure; that is, momentum input is not significant.

The parameterisation of the intrusion of the inflowing river at the level of neutral buoyancy as parameterised by Imberger et al (1976) is incorporated in DYRESM in terms of the parameter, $R_i$, where

$$ R_i \sim F_i Gr^{1/3} $$

(6)

where the Grashof number $Gr \sim N^2L^4/K_i^2$, and $L$ is the reservoir length at the
level of insertion. For $R_i < 1$, the intrusion is governed by a viscous-buoyancy balance, for $R_i > 1$, by an inertia-buoyancy balance.

In the Wellington Reservoir inflow studied by Hebbert et al (1979), $N^2 \sim O(1.8 \times 10^{-3} \text{sec}^{-2})$, and the length of the reservoir is $O(20 \times 10^3 \text{m})$. Using $10^{-5} \text{m}^2 \text{s}^{-1}$ as a typical $K_z$ yields $Gr \sim 3 \times 10^{24}$, and with $F_l \sim 0.2$, $\frac{R_i}{\theta} \sim 107 >> 1$. In this period of high inflow, the intrusions are clearly inertia dominated. In periods of low flow, however, $R_i < 1$ and the intrusions are governed by a viscous buoyancy balance. In Kootenay Lake, the July 1976 inflows yields $N^2 \sim 5 \times 10^{-8} \text{sec}^{-2}$. With a lake length $O(100 \times 10^3 \text{m})$ and $K_z \sim 10^{-5} \text{m}^2 \text{s}^{-1}$, $Gr \sim 5 \times 10^{25}$, and $R_i \sim O(10^8)$. Again the major inflow intrusions are inertia dominated during this period; in low flow periods, $R_i < 1$, and viscous intrusions are.

In both lakes the inflow intrusions may then fall in either the viscous or inertia dominated regimes; the plunging of the inflow and entrainment from the body of the lake may however be expected to be of greater significance in the Wellington Reservoir, whereas cabling will play an important role in Kootenay Lake. In both cases, the $F_l$ values are sufficiently small to ensure that the one-dimensional parameterisation described by Imberger et al (1976) will be applicable. The influence of Coriolis forces on the inflow in each reservoir is small; in both cases the Rossby radius based on the density difference between inflow and surface water and the flowing depth is large in comparison to the scale of the lake width. Finally, there is a basic difference in the types of inflow. Wellington Reservoir inflows are unregulated, and clearly follow the seasonal pattern (Figure 2); on the other hand, the inflows to Kootenay Lake are regulated by upstream storages; with a consequent smoothing of the large inflow peaks (Figure 5) (Daley et al 1980).
Outflow

The outflows from the two reservoirs are also significantly different. The Wellington Reservoir withdrawal is taken from one or both of two submerged outlets at the base and 15 m above the base of the dam wall, the draw being manipulated according to demand and operating constraints. On the other hand, the outflow from Kootenay Lake is uncontrolled flow over a submerged sill.

A number of possible modes of withdrawal from these outlets are possible. Firstly, if the stratification is very weak, the flow into the sink will be essentially potential, and all levels of the storage will be sampled. Secondly, as the stratification increases, buoyancy forces restrict the vertical motions and the withdrawal is confined to a narrow layer at the level of the offtake. Thirdly, if the stratification is such that the withdrawal layer intersects the pycnocline, the pycnocline may be drawn down or up into the sink and water from both above and below the pycnocline will be sampled.

These problems have recently been reviewed by Lawrence and Imberger (1979) and Imberger (1980) and will not be discussed here except to note that a critical Froude number, $F_c$, may be found for which drawdown or drawup of a pycnocline will occur; for a sink at the bottom of a semi-infinite reservoir, $F_c = 1.27$, for a sink away from the bottom $F_c = 2.54$, where

$$F = \frac{Q}{(g'd^5)^{1/2}},$$

(7)

where $Q$ is the discharge, $g'$ the reduced gravity across the pycnocline, and $d$ the distance between sink and pycnocline. Thus for a given $Q$ and...
stratification, a critical value of $d$ may be found; if $d_c > d$, it may be anticipated that drawdown will occur.

To obtain a measure of the likelihood of this occurring in Wellington Reservoir, the equivalent two layer structure generated in the previous section may be used to find $g'$ and $d$, while $Q$ is available from the data. The results of this computation indicate that, in the Wellington Reservoir, drawdown occurs only infrequently and the two and three-dimensional theories described by Imberger et al (1976) and Lawrence and Imberger (1979) will apply. Drawdown only occurs at those times when the pycnocline is close to the operating offtake.

These criteria are based on a point sink located in an infinite fluid. While there is some justification for their application in Wellington Reservoir, where the sinks face the length of the reservoir, their applicability in Kootenay Lake is somewhat more doubtful. In this case, the sink is effectively a vertical distribution of line sinks parallel to the length of the lake, facing only the lake width of $O(3 \text{ km})$. There is apparently no theory available to cover this situation, and in the absence of other criteria, those above are applied as an indicator.

In this case there are strong indications that drawup (in this case) of the pycnocline into the outflow over the sill will occur. On all but a few occasions the critical value is exceeded, and on these occasions the pycnocline is at least 50 m removed from the sill. It might therefore be expected from this theory that water from both above and below the pycnocline will comprise the outflow from Kootenay Lake, according to the ratio given by Wood (1978).
For the case that drawdown does not occur, the withdrawal layer evolution depends on the value of the parameter

$$R_3 = \frac{Q}{\nu L},$$

(8)

where $\nu$ is the kinematic viscosity and $L$ the reservoir length (Lawrence and Imberger 1979). For $R_3 > 1$, the three-dimensional layer thickness is governed by an inertia-buoyancy balance, and its thickness is given by

$$\delta_3 \sim O\left(\frac{Q}{N}\right)^{1/3}$$

(9)

where $N$ is the buoyancy frequency. For $R_3 < 1$, the layer is the result of a viscous buoyancy balance and

$$\delta \sim O\left(\frac{L}{N}\right)^{1/3}$$

(10)

as for the equivalent two-dimensional theory (Imberger et al, 1976).

For the Wellington Reservoir, $L \sim O(30 \text{ km})$ and using eddy values of $\nu$, $R_3 > 1$ gives $Q > O(3 \text{ m}^3\text{s}^{-1})$. This is a frequent occurrence during the simulation period. In Kootenay Lake, as the sill lies along the axis of the lake $L \sim O(1 \text{ km})$, and $R_3 > 1$ yields $Q > O(0.1 \text{ m}^3\text{s}^{-1})$ a situation which is invariably the case. Thus in the Wellington Reservoir either case may arise and the thickness of the withdrawal layer will be given by the appropriate one of $\delta$ and $\delta_3$ above; in Kootenay Lake, the withdrawal will always be governed by an inertia buoyancy balance and the layer thickness estimated by $\delta_3$.

The remaining point of interest with respect to withdrawal is its
influence on the predominantly one-dimensional structure. This is best judged by considering the value of the Froude number, $F_0$, where

$$F_0 = \frac{Q}{(g'H^2)^{1/2}}$$

(11)

where $g'$ the reduced gravity over the total depth $H$. If $F_0 < 1$, the withdrawal will not disturb the one-dimensionality by introducing significant vertical motions.

In the Wellington Reservoir, $F_0$ is generally $O(10^{-2})$, and even during the peak withdrawal periods, does not approach $O(1)$. The same is true for Kootenay Lake.

In general therefore the indications are that the withdrawals from the two lakes are of different character, with the Wellington Reservoir outflows being predominantly selective withdrawal from a specific region of the reservoir, whereas in Kootenay Lake the draw will frequently sample different regions simultaneously. The differences in elevation and control of the withdrawals will also have a marked effect on the resulting stratification. For example, the Wellington Reservoir draws are frequently taken from below the pycnocline, with the result that the structure is weakened from below. Draws from above the pycnocline will have the effect of sharpening the interface. On the other hand, the withdrawal from Kootenay Lake is generally above the pycnocline level, but the draw up phenomena will also sample from below the interface, effectively smearing the interface vertically.
4. SIMULATION RESULTS

The Simulation Model

The model DYRESM is a one-dimensional model for the prediction of temperature and salinity profiles in lakes and reservoirs of small to medium size. As indicated in section 1, the size limitation on the model is one of applicability of the assumption of one-dimensionality. As the model is based on a parameterisation of the dynamics of the processes active in the lake, it is possible to quantify the validity of the assumption in terms of the non-dimensional numbers \( W, F_0, F_1, \) and \( R \) as defined in the previous section. As shown above, both Wellington Reservoir and Kootenay Lake may be classed as one-dimensional in a mean sense, with the latter near to the limit of applicability.

The concepts and architecture of the model and the process parameterisations have been fully described in Imberger and Patterson (1981), and only brief details will be given here.

Basically the model concentrates on the parameterisation of the physical processes rather than solution of the appropriate differential equations and makes use of a Lagrangian layer concept, in which the reservoir or lake is modelled as a system of horizontal layers of uniform property. These layers may move vertically, changing thickness in accordance with the given depth-volume relationship, as the processes of inflow and withdrawal change the total volume. The result of this concept is a model which is simple, computationally and conceptually realistic in the one-dimensional sense. Because of the simplicity, the algorithm is economical, and long term simulations which incorporate the finest scales become possible.
There are five basic physical models incorporated in DYRESM; surface heat exchanges, mixed layer dynamics, inflow and outflow dynamics, and vertical diffusion in the hypolimnion. The parameterisations of each of these processes and the associated computational strategy is treated in complete detail in Imberger and Patterson (1981) and will not be repeated here.

Wellington Reservoir

The simulation of the reservoir was initialised by the mean field data profile taken on May 13, 1975 (day 75133) and allowed to run without adjustment for a 964 day period to December 31, 1977 (day 77365), the end of the input data set. Daily inflow, outflow, and meteorological data provided model input. The mean field profiles served as a basis of comparison of the model prediction with reality. The results of the simulation are shown in Figure 7, 9 and 10.

The predictions made by the model may be examined on two time scales; seasonal and daily. On the seasonal scale, which is of most interest to water quality concerns, the isotherm and isohaline positions generated by the model are plotted as a function of time in Figures 9a and 9b. Comparison of the isotherm plots 9a and 3a show that the seasonal thermal behaviour is reproduced well, with all the mixing and stratifying events occurring at the right times and of the correct magnitudes, noting that the temporal resolution of the model is much finer than that of the field data. The principal anomaly is the appearance of the deep cold (12°C) water in August 1975. This is not present in the field data and results from the inflows. Small errors in the inflow temperature data in this period are the most likely cause of this mismatch. In any case the predicted temperature gradient in the base of the reservoir at this time is small and
the actual discrepancy between field and predicted temperatures is $0(0.5^\circ\text{C})$. There are other minor anomalies present, but in general the agreement on the seasonal scale is good.

The observed and predicted isohaline positions (Figures 3b and 9b) also compare well, with the most significant discrepancy being during the 1976 winter, where the model has evidently supplied insufficient mixing. As shown on Figure 7a, the parameter $W$ changes rapidly through several orders of magnitude in this period, a time of highly fluctuating wind speeds and weak stratification. Small errors in the wind speed data (taken from a meteorological station some distance from the reservoir) could account for this discrepancy. The model does not reproduce the peak salinities in the base of the reservoir during the winter inflow period. This is principally due to the layer structure of the model, the layer thicknesses and hence the vertical resolution being determined in part by volume rather than height resolution constraints. This results in the bottom few metres always appearing as mixed, and the error is one of display rather than salt content and is accepted in the interests of economy.

It is also of interest to compute the predicted values of the parameters $W$, $R$ and $F_i$ for the simulation period. This is done the same way as previously using the predicted density profile rather than the mean field profile. The parameter values, computed at five day intervals, are shown as the solid lines on Figure 7. The extreme variability of $W$ is clearly shown and would likely increase were single day intervals taken. However, the field based and predicted values generally compare well, indicating that the mixed layer depths correspond, as well as the period of internal seiching.
The predicted Rossby radii also track the field based values well, indicating that the average density gradients are reproduced well by the model. The predicted and field based inflow Froude numbers are also substantially in agreement, suggesting that the density gradients in the region of inflow insertion are properly represented by the model.

Comparison of W, R and F\textsubscript{1} from both field and predicted profiles gives a measure of the accuracy of the reproduction of the vertical structure over the simulation period; the timing and magnitude of events is tested by comparison of Figures 3 and 9. A quantitative measure of absolute fit is also required.

Volume weighted temperature averages are computed from both model and field profiles as

\[
\bar{T}_m = \frac{1}{n} \sum_{i=1}^{n} \frac{V_i T_{m_i}}{\sum_{i=1}^{n} V_i}, \quad \bar{T}_f = \frac{1}{n} \sum_{i=1}^{n} \frac{V_i T_{f_i}}{\sum_{i=1}^{n} V_i}
\]  

(12)

where the subscripts \text{m} and \text{f} refer to model and field values respectively, the subscript \text{i} to the value of the parameter at height \text{z}_i from the deepest point, and \text{V}_i is the volume contained between heights \text{z}_i and \text{z}_{i-1}. The root mean square error \text{E}_1 and the volume weighted root mean square error \text{E}_2 are calculated from

\[
E_1(T) = \sqrt{\frac{1}{n} \sum_{i=1}^{n} (\bar{T}_{m_i} - \bar{T}_{f_i})^2}
\]

(13)
Similar formulations apply for the corresponding salinity parameters $S_m$, $S_f$, $E_1(S)$, $E_2(S)$.

It needs to be noted in the interpretation of these parameters that the resolution of the model output is much finer than that of the field data. The summations above are taken over the model output heights, with interpolated field data at those heights. Thus structure with a scale less than the field data resolution will not be present in the $T_f$ and will therefore register as an error which is not present. On the other hand, interpolating onto the field data heights yields large errors in the weighted parameters as the result of the large incremental volumes associated with the relatively widely separated heights. The parameters used here are calculated by the first method.

The values of the parameters for representative days from the available Wellington Reservoir field data are shown in Table 1. The averages of the parameters are taken over all possible field data days.

A comparison of $T_m$ with $T_f$ and $S_m$ and $S_f$ yields an indication of the ability of the model to correctly predict the heat and salt content of the reservoir. Table 1 shows that the heat content predicted is, on average, with 2% of that estimated from the field data, whereas the salt content is within 20% of the estimated value. In view of the resolution problem noted above, the comparisons are considered satisfactory.
<table>
<thead>
<tr>
<th>Day</th>
<th>$T_m$</th>
<th>$T_f$</th>
<th>$E_1(T)$</th>
<th>$E_2(T)$</th>
<th>$\bar{T}_m$</th>
<th>$\bar{T}_f$</th>
<th>$E_1(S)$</th>
<th>$E_2(S)$</th>
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<td>15.605</td>
<td>0.115</td>
<td>0.644</td>
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<td>314.7</td>
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<td>0.459</td>
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<td>0.236</td>
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<td>725.6</td>
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<td>22.402</td>
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<td>530.3</td>
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<td>48.7</td>
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<td>Average (overall data)</td>
<td>16.466</td>
<td>16.123</td>
<td>0.140</td>
<td>0.880</td>
<td>520.7</td>
<td>639.7</td>
<td>24.7</td>
<td>150.3</td>
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</tbody>
</table>

Wellington Reservoir simulation performance statistics
The unweighted parameters $E_1(T)$ and $E_1(S)$ may be used to give an estimate of the absolute fit of the predicted profile to the field profile, and may be interpreted as expected errors. Thus, on average, the predicted profile is within 0.14°C and 25 ppm of the field profile. These indicators of course tend to bias the performance in the hypolimnion where the $V_t$ are small and the dynamics are restricted to inflow, outflow and internal turbulent mixing. On the other hand, $E_2(T)$ and $E_2(S)$ bias the result in the epilimnion, where the $V_t$ are large and the mixed layer dynamics predominates. In this case, the epilimnion temperatures and salinities predicted are on average within 0.9°C and 150 ppm of the estimated field values. It is the epilimnion, however, where structure of a scale less than that of the field data resolution occurs, and as noted above, a false indication of error occurs. The values of $E_2(T)$ and $E_2(S)$ are therefore considered artificially high.

The second level of comparison for the model is at the daily time scale. Figure 10 shows the predicted temperature (solid line) and salinity (broken line) profiles, together with the mean field temperature (circles) and salinity (triangles) profiles for three days in the simulation period. The days were chosen on the basis of being roughly equally spaced and are typical of the fit obtained throughout the simulation period. The days represented are 76040, 77013, and 77320, respectively 273, 612 and 919 days from initialisation. There are clearly some discrepancies present, however, even after 919 days the structure is accurately represented and the temperature and salinity predictions are within 1.5°C and 200 ppm. The absence of the large salinity spike in the predicted profile on day 77320 is the result of either faulty inflow salinity data or incorrect field profile data.
Kootenay Lake

The simulation of the thermal and salinity structure of Kootenay Lake was initialised by the whole-lake average profile on June 16, 1976, (day 76168) and was allowed to run continuously until November 16, 1977 (day 7730). The results of simulation by DYRESM for this 519 day period are displayed in Figures, 8, 11 and 12.

Again the predictions made by the model may be examined on both seasonal and daily time scales. On the seasonal scale, the predicted isotherm and isohaline depths as a function of time are shown in Figures 11a and 11b, to be compared with Figure 6a and 6b respectively.

The comparison of the observed and predicted thermal behaviour, Figures 6a and 11a, shows that, although there are some marked differences, the seasonal variation of the thermal structure occurs at the right times and of the correct magnitude. As mentioned above, the resolution of the field data is such that the representation of mixing events, for example in Autumn 1976, is spread temporally by the interpolation, and the model representation, which has a much higher resolution, actually fits the field data quite well. The same may be said of the restratifying and subsequent mixing the following year. The principal departure in the summer periods is the degree of stratification; clearly the model has produced considerably sharper thermoclines through both summer periods. This may be due to the smearing effect inherent in the averaging of the field profiles, as mentioned above. It may also be due to a two-dimensional effect not correctly parameterised by the model. For example, with internal seiche periods of several days, it is quite possible that differential deepening of the tilted thermocline has occurred. An additional possibility is that the seiching has generated local mixing on
the lake boundaries, with the resulting homogeneous fluid penetrating horizontally, spreading the interface (Imberger and Hamblin, 1982). These effects are not included in the model. A third possibility is that the departure is the result of incorrect parameterisation of the withdrawal. As shown in section 3, there are strong indications that drawup of the pycnocline into the overflow over the sill will occur. Although based on a theory not strictly applicable to the Kootenay Lake case, these indications suggest that the two-dimensional selective withdrawal algorithm incorporated in the model may lead to error. In particular, the withdrawal layer computed by the existing algorithm will usually encompass the uniform mixed layer, but its extent will be limited by the interface. Thus the withdrawal will take only fluid from above the interface, leading to a sharpening of the structure at the pycnocline. On the other hand, if drawup of the pycnocline occurs, fluid from both above and below the interface will be removed, effectively weakening the structure. This is clearly a serious deficiency in the model, and incorporation of a parameterisation of drawdown is currently in progress.

During the winter period, the model has evidently produced insufficient mixing and the temperature inversion is apparently very much weaker than that indicated on Figure 6a. However, in view of the comments made earlier with respect to resolution, and the fact that the actual temperature difference between predicted and observed profiles is very small, the error here is in fact small. The prediction of the winter inversion will be discussed in more detail below.

The comparison of the isohaline positions (Figures 6b and 11b) on a seasonal scale yields the same kind of result. During the summer period the variations is of the right order, occurring at the correct times.
Again, there appears to be insufficient mixing in early winter. The rapid change in the position of the 140 ppm and 150 ppm isohalines in early March 1977 is due to an isolated wind event on day 77066.

The computed values of $W$, $R$, and $F_i$ are shown as the solid lines on Figure 8. The extreme variability of $W$ is again evident, however, the field based and predicted values compare well, indicating the success of the mixed layer algorithm. The predicted Rossby radii and inflow Froude numbers also agree with the field based values.

For a quantitative measure of the overall success of the model, the parameters $T_m$, $T_f$, $E_1(T)$, $E_2(T)$, $S_m$, $S_f$, $E_2(S)$, and $E_2(S)$ are again calculated from (12) and (13) for all available field data days. The results are shown in Table 2. The predicted heat and salt contents, as indicated by $T_m$, $S_m$ are within 7% and 1% respectively of the field predicted values. The measures of absolute fit of the profiles, $E_1$ and $E_2$, indicate that the expected error in temperature less than, on average, 0.2°C and in salinity 1 ppm. The weighted indicators give expected errors in the epilimnion of within 1°C and 4 ppm respectively. The rapid increase in $E_1(T)$ and $E_2(T)$ on day 77147 is due to a poor fit around the thermocline. In particular, the predicted thermocline is much sharper than that represented by the field data, which shows no distinct mixed layer. Consequently the model result is deficient in heat below the level of the mixed layer. The temperature in the epilimnion is largely controlled during the following summer by surface heat transfers and therefore adjusts itself to the correct value. The heat content of the hypolimnion, however, is largely insulated from the epilimnion by the mixed layer, and this heat deficiency will there be retained by the model until the lake next becomes isothermal.
<table>
<thead>
<tr>
<th>Day</th>
<th>$T_m$</th>
<th>$T_f$</th>
<th>$E_1(T)$</th>
<th>$E_2(T)$</th>
<th>$\overline{T}_m$</th>
<th>$\overline{T}_f$</th>
<th>$E_1(S)$</th>
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<td>135.6</td>
<td>135.0</td>
<td>0.5</td>
<td>4.0</td>
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</table>
The reason for this anomaly at the beginning of the summer stratification period is not clear; any of the mechanisms suggested above could be responsible. However, it is plausible to suggest that since a significant mixing event has occurred to form a mixed layer, boundary mixing generated by internal seiching is the cause of the interface smearing.

On the daily scale, the comparison of three typical days is shown in Figure 12. The days chosen include 76196, 77058 and 77320, being representative of the summer, winter, and following summer periods respectively. On day 76196, 38 days after the start of the simulation, two factors are immediately evident. Firstly, the model has predicted a shallow mixed layer which is not evident on the field data. Corresponding to this is a mismatch in temperature above and below the predicted thermocline. The total heat content is, however, accurately predicted. Secondly, there is a significant salinity anomaly in this mixed layer region, the base of which corresponds to the depths at which the major rivers penetrate. This lack of agreement may be due to errors in the inflowing salinity data caused by uncertainty in the interpolation of salinity inflow over monthly intervals and by the contribution of ungauged inflow to the salinity balance, particularly in June, 1976. It should also be noted that the vertical resolution of DYREMS is several times greater than that of the field observations.

On day 77058, 256 days after the start of the simulation, a typical winter profile has appeared. The lake is virtually isothermal, with the salinity content of the hypolimnion being sufficient to arrest deepening of the mixed layer. There is a weak temperature inversion at an elevation of approximately 80 m, as reported by Wiegand and Carmack (1981). The only significant anomaly in this comparison is the sharpness of the
salinity gradient at the base of the mixed layer.

The third daily comparison is made on day 77320, the end of the simulation period. The only significant anomaly present is the heat deficit below the predicted mixed layer, which has propagated through from day 77147, as referred to above. This deficit is present in all predicted temperature profiles beyond 77147.

It is of interest to examine the winter temperature inversion in more detail. Unfortunately only two whole lake surveys were carried out during the period of the existence of the inversion, those on days 77026 and 77058. However, the fixed temperature profiler located at Pilot Bay in the mid-lake region (see Figure 4) provided data through this period. These data were presented by Wiegand and Carmack (1981) in 24 hour average form and are reproduced here as Figure 13a. These profiles show the presence of the inversion for the first time on January 14, 1977 (day 77014) centred at an elevation of about 75 m. The inversion then persisted through winter, contained between elevations of roughly 40 m and 100 m. Its thickness decreased through March, and by the end of March, the inversion had disappeared. Wiegand and Carmack (1981) describe the formation of the inversion as the result of surface driven mixing processes being insufficient to penetrate the halocline which formed at an elevation of about 80 m; the temperature inversion is therefore the remnant of the previous summer's stratification.

With the aid of the simulation result it is possible to analyse in detail the formation and erosion of the inversion. In Figure 13b are shown the predicted temperature profiles over the same period illustrated in Figure 13a. In making an absolute comparison of Figures 13a and 13b
it should be noted that the field profiles are representative of the mid-lake region whereas the model predictions are of course whole lake mean results.

The inversion predicted by the model comes into existence some ten days after the time indicated by the fixed temperature profiler and disappears some twenty days earlier. In all other respects it is qualitatively very similar; the depth and size of the inversion are of the same order. The model has produced a sharper inversion than indicated in Figure 13a, consistent with the sharper thermocline predicted throughout the simulation. Of more interest, however, are the formation and erosion of the inversion.

The formation of the inversion occurs roughly as described by Wiegand and Carmack (1981). The stratification present on day 77002 is eroded by surface cooling and wind stirring, generating a mixed layer some 70 m deep by day 77005. The rate of descent of the mixed layer decreases as the mixed layer temperature approaches the temperature of maximum density of \( \frac{\partial p}{\partial T} \) becomes small. In the model simulations, \( \frac{\partial p}{\partial T} \) always remains negative. The cooling and deepening continues at a reduced rate until, on day 77022, the mixed layer is some 80 m deep. A haline has also formed at this depth as the result of this deepening, the salinity gradient being 0(2.9 ppm/m). The mixed layer temperature and salinity are 4.39°C and 132 ppm respectively, and the corresponding density 1000.077 kg m\(^{-3}\) (Chen and Millero, 1977). The density gradient directly below the mixed layer is 2.39 x 10\(^{-3}\) kg m\(^{-4}\). Deepening by free convection can only occur as long as the buoyancy flux is negative, that is while the mixed layer temperature is cooling towards the temperature of maximum density \( T_{md} \) (in this case \( T_{md} = 3.9545°C \) for a salinity of 132 ppm);
thus even more the mixed layer temperature to drop to $T_{md}$, the mixed layer density would increase to 1000.0875 kg m$^{-3}$, resulting in a deepening of 0.6 m, ignoring further depression of $T_{md}$ by the increased salinity of the mixed layer. As the halocline is somewhat thicker than this, it is therefore impossible for surface cooling to generate further deepening. Only wind stirring or wind induced shear at the interface could further influence the mixed layer depths; both of these effects are negligible at the time of formation.

The inversion is therefore the result of further surface cooling. As $\partial \rho / \partial T$ is small, the density changes are also small and the inversion develops as the result of the restricted mixing, supported by the relatively strong halocline. The river inflows appear to have no influence on the generation of the inversion, with the Kootenay River plunging to well below the inversion level and the Duncan River overflowing as a surface layer. The Duncan River inflow is responsible for the small temperature reversal at the surface; the inflow is cold (0°C) and relatively fresh (0(100 ppm)) and is therefore relatively buoyant.

The subsequent erosion of the inversion is also of interest. Wiegand and Carmack (1981) characterise the erosion by an eddy diffusion process, and evaluate a diffusion coefficient of $1.5 \times 10^{-4}$ m$^2$s$^{-1}$ for the period, an order of magnitude larger than the maximum value suggested by the simulation. Other processes are, however, certainly relevant. For example, the sharpening of the inversion on March 6 (day 77065) indicated by the fixed temperature profiler on Figure 13a can only be due to a Kootenay River inflow entering at the inversion level, as there has been an increase in temperature over the whole profile. The increases above and below the inversion may be due to eddy diffusion and, in the upper
region, surface heating, whereas the inversion itself can only have received a heat input from a warm, saline inflow.

The model profiles indicate that the inversion is no longer present on day 77066 (March 7); prior to this surface heating and subsequent cooling and weak wind events had generated a multiple thermocline structure. Although weakened by turbulent diffusion processes, the inversion is still present on day 77065, with a thermocline located some 20 m above. On day 77066, however, the average wind speed over the lake rose to 6.8 ms\(^{-1}\), and the subsequent deepening of the thermocline due to both surface stirring and shear production was sufficient to penetrate the inversion which was absorbed into the mixed layer. Thus the ultimate demise of the inversion was the result of a single wind event following a gradual weakening.

The strengthening of the inversion on March 6 by the inflow suggested above is not evident in the model predictions. However, the Kootenay River inflow is modelled as penetrating at about the level of the inversion but having little effect on its strength. As the inversion is at this stage of \(0.03^\circ C\) and the level of insertion of the inflow largely controlled by salinity, a small error in the inflow temperature could account for the lack of effect. A further, deeper inversion is predicted later in the month and is directly attributed to the Kootenay River inflow. This inversion appears transiently on day 77072 and again on 77081, in each case surviving for a few days before another slightly warmer inflow is inserted at the level of the temperature decrease. On day 77088 another inversion is generated in the same way and is supported and strengthened by further warm saline inflows at a slowly increasing level until days 77095 when the inflow insertion level rises well above the inversion, which subsequently simply diffuses away, still being weakly present some
60 days later. These later inversions are not indicated by the mid-lake data (Figure 13a); however, in this context it should be noted that inflows, which in fact may take several days to penetrate to the central region of the lake, are completely inserted in one day by the model. Thus the model results may anticipate the mean field result by several days.
5. DISCUSSION

In the foregoing the simulation of each of the water bodies has been described. Although the reservoir and the lake have aspect ratio (length/total depth) of the same order, the much greater depth of Kootenay Lake is associated with a much larger hypolimnion depth. As a result, the temperature distribution in Kootenay Lake is determined less by the inflows and meteorological forcing and more by the vertical mixing processes than is the case in the Wellington Reservoir. Thus Kootenay Lake has a much longer memory of forcing events than does Wellington Reservoir, which is very much locally forced. Clearly the dynamics of the two water bodies are very different; the mixed layer dynamics of Kootenay Lake being dominated by shear production and billowing at the interface as the result of internal seiching whereas the Wellington Reservoir mixed layer is generated by surface stirring; the withdrawal from Kootenay Lake is at the surface level whereas in the reservoir the withdrawals are primarily subsurface; the Kootenay Lake inflows rarely replenish bottom water in contrast to the reservoir situation where most of the winter inflows underflow the stored water. In this context, the two water bodies represent extremes of behaviour and are therefore a good test of the capabilities of the simulation model.

In the transferral of the model from Wellington Reservoir to Kootenay Lake only minimal changes were made. The most important of these was concerned with the deepening of the mixed layer by shear production, expected to be dominant in the lake. As mentioned briefly above, the internal seiche period of Wellington Reservoir is generally $O(10$ hours), frequently less than the duration of a wind event, whereas in Kootenay Lake, as a consequence of the much greater length, the internal seiche
period is 0(7 days), much greater than the duration of a wind event.

As detailed in Section 3, the parameterisation of shear production is based on the work of Spigel and Imberger (1980) and relies on the wind event remaining constant for at least one quarter period of the internal seiche. This is usually the case in the Wellington Reservoir, but rarely so in Kootenay Lake. According to Spigel and Imberger (1980) shear production at the interface is cut off after one quarter period as the influence of the boundaries is felt; in the context of Kootenay Reservoir, this parameterisation was modified to incorporate changes in wind direction at six hourly intervals. Thus in the Kootenay Lake simulation, shear production is cut off after one quarter period has elapsed or if the component of six hourly averaged wind along the lake changes sign, whichever occurs earliest (generally the latter).

Apart from this alteration the model changes were limited to the physical description of the water body, or to different data inputs. For example, directly measured incident long wave radiation was available for Kootenay Lake, whereas this quantity was computed from an empirical formula in the reservoir simulation (Imberger and Hebbert, 1980). In addition, the availability of measured extinction coefficients for short-wave radiation in Kootenay Lake allowed the use of a time dependent extinction coefficient $\eta$ of the form

$$\eta = 0.15 \cos\left(2\pi \frac{(D-165)}{365}\right) + 0.35 \text{ m}^{-1} \quad (14)$$

where $D$ is the day number within a particular year.

The level of withdrawal in the Kootenay Lake simulation was set at
the surface level rather than at the sill level in an attempt to achieve a more realistic horizontal velocity profile from the two-dimensional theory, and hence a more appropriate distribution of withdrawal volumes. The parameterisation of withdrawal was not changed.

The importance of vertical mixing is emphasized by the Kootenay Lake simulation. The overall parameterisation of vertical exchange is that of a three component system: in the upper region mixing is determined by wind stirring, convective motions set up by surface cooling, and internal wave induced shear and billowing at the interface all of which are represented in the model by the mixed layer parameterisation. Below the mixed layer a region exists where the vertical diffusivity is inversely related to the density gradient, the relationship depending on the instantaneous energy input of wind and inflows. In the remaining portion of the water column, generally near the bottom, where stratification is weak the diffusivity is assumed to be constant at a prescribed limiting value (see Imberger and Patterson, 1981). Lam and Schertzer (1980) reached a similar formulation from their seasonal simulations of the thermal structure of Lake Erie.

The two parameters involved in the formulation of the eddy diffusivity are a constant of proportionality and the limiting value. The constant may be related to the efficiency of the mixing (Imberger and Patterson, 1981 and Imberger and Hamblin, 1982) and has remained constant through the simulations. The limiting value of the diffusivity has also remained constant (at $10^{-4}$ m$^2$s$^{-1}$). The results of a simulation, however, are quite sensitive to the limiting value, as evidenced by Figure 14 which shows the profile on day 77320 resulting from a simulation using a limiting value of $10^{-3}$ m$^2$s$^{-1}$, a tenfold increase. When compared with the
appropriate profile of Figure 12, a strong discrepancy has been introduced. It is interesting that a limiting value of $10^{-5} \text{m}^2\text{s}^{-1}$ proved optimum for both simulations, although the Wellington Reservoir simulations were less sensitive to changes than those of Kootenay Lake.

The limiting value of the eddy diffusivity remains as the only calibratable parameter in the model and as such is the principal shortcoming of the model. A recent paper (Weinstock, 1981) addresses this problem and experimental modifications to the algorithm incorporating Weinstock's result have had promising results.
6. CONCLUSIONS

The emphasis in this paper has been placed on the evaluation of a model based on the parameterisation of the physical processes most affecting the vertical distribution of heat and a material substance in two water bodies at extreme ends of the range of applicability of the model. No changes to parameter values were made in the transference from a reservoir to a lake. The results on both water bodies were good, indicating the accuracy of the parameterisations, although clearly some two and three-dimensional effects are beginning to influence the results in Kootenay Lake.

This approach is in contrast to that of attempting a best fit set of model coefficients on a case by case basis, which inherently does little to aid the understanding of the physical processes involved. The physical approach taken here gives the user confidence in the result, and allows an understanding of individual phenomena to be gained; for example, the formation and erosion of the Kootenay Lake winter temperature inversion may be analysed in detail from the simulation results.

It is also appropriate to mention here that in the preparation of the simulations a number of errors in the wind data were discovered, in both cases, through a lack of agreement between the observed data and the early simulation results. In both cases overmixing occurred to varying degrees; these were traced through the details of the mixing parameterisation to an apparently excessive wind input. A check of the data confirmed processing errors in both cases. A similar error in inflow salinity in Wellington Reservoir was uncovered in the same way. Thus the simulation model may be used to test for consistency between physical data sets in lakes and reservoirs.
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REFERENCES


LEGENDS TO FIGURES

Figure 1  A map of Wellington Reservoir, showing the location of the physical monitor stations.

Figure 2  Wind speed, solar radiation, inflow salinity, inflow temperature, and inflow flux for Wellington Reservoir for the simulation period.

Figure 3  The length averaged (a) temperature and (b) salinity structure as a function of time for Wellington Reservoir, from the field data.

Figure 4  A map of Kootenay Lake, showing the location of the physical monitor stations, the radiation station, and the meteorological buoys.

Figure 5  Wind speed, solar radiation, inflow salinities, inflow temperatures, and North Arm inflow flux for Kootenay Lake for the simulation period.

Figure 6  The length averaged (a) temperature and (b) salinity structure as a function of time for Kootenay Lake, from the field data.

Figure 7  Computed parameter values for Wellington Reservoir from the field data (symbols) and model prediction (solid line); (a) Wedderburn number $W$; (b) Rossby radius of deformation, based on the mixed layer scale; (c) Inflow Froude number.

Figure 8  Computed parameter values for Kootenay Lake, as for Figure 7.

Figure 9  The model predictions of (a) thermal, and (b) salinity structure for Wellington Reservoir.

Figure 10  Typical observed and predicted mean profiles for Wellington Reservoir.

Figure 11  The model predictions of (a) thermal, and (b) salinity structure for Kootenay Lake.
Figure 12  Typical observed and predicted mean profiles for Kootenay Lake.

Figure 13  (a) Details of the temperature inversion observed by the mid-lake fixed temperature profiler in Kootenay Lake (reproduced from Wiegand and Carmack, 1981, with permission of the authors), starting at day 77002, with two day increments.

(b) The predicted temperature inversion, corresponding to the same days in (a).

Figure 14  The observed and predicted profiles on day 77320 in Kootenay Lake with an increase in the maximum eddy diffusivity to $10^{-5}$ m$^2$s$^{-1}$. 
Fig 3
Fig 4

• Physical monitor station
• Radiation station
• Meteorological buoy

SCALE

0 km 6 km

CONTINUED ON B

N17
N16
N15

PILOT BAY F.T.P.

N1
N2
N3
N4
N5
N6
N7
N8
N9
N10
N11
N12
N13
N14

SCALE

0 km 5 km

CONTINUED ON B

117.00W
Fig 10
Fig 13