Master's Thesis:

Numerical Modelling of Stratification in Lake Constance
with the 1-D hydrodynamic model DYRESM

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Tasks and Objectives
Data collection and preparation

Investigation whether DYRESM can be the hydrodynamic component of an ecosystem model for Lake Constance

Data collection and preparation
• Forcing data (meteorological data, inflow and outflow data)
• Geometrical data
• Field data (temperature & salinity profiles, water levels)

DYRESM standard version
• thermocline too sharp
• surface water temperatures too high

DYRESM extended version

Extended DYRESM inclusion of turbulent diffusion

Extended DYRESM

Calibration

Optimal parameters set:
C = 1500
WMF = 1.45
PLT \text{max} = 2.5 \text{ m}

Other parameters held constant in calibration
• Mean light extinction coefficient $k = 0.35 \text{ m}^{-1}$
• Minimum permissible layer thickness $PLT_{\text{min}} = 0.5 \text{ m}$

Sensitivity analyses
Low sensitivity to $PLT_{\text{min}}$
Medium sensitivity to $k$
High sensitivity to $\theta$ and $CCF$

Verification

The extended version of DYRESM achieves after calibration good simulation results for Lake Constance.
The coupling to the aquatic ecosystem model CAEDYM is recommended for future model applications on Lake Constance.
DYRESM indicates a strong response of Lake Constance to simplified climate change scenarios.
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Chapter 1

Introduction

The first section of this chapter explains the goals of this thesis (1.1), whereas the second section provides information on the lake of study, Lake Constance (1.2). The third and last section discusses the tool that was used, the numerical model DYRESM (1.3).

1.1 Objectives of this Thesis

Lake Constance, one of Europe’s largest lakes, is located on the northern edge of the Alps, shared by Germany, Switzerland and Austria. The good water quality, the close vicinity to major cities and its beautiful setting along the Alps may be reasons for Lake Constance’s popularity for vacation, recreation and water sports. Besides tourism, which contributes significantly to the local economy, a small fishing industry makes a living from the lake’s natural resources.

Of extraordinary relevance however is, that the lake supplies 165–185 million \( m^3 \) of high-quality drinking water each year, enough for 3.5–4.5 million people (LfU, 1997). The majority of the withdrawn water, about 130–140 million \( m^3/yr \), is distributed in the greater region of Stuttgart, some 200 km north of Lake Constance. Although surface waters are generally more vulnerable to pollution and spills than groundwater, the chemical composition of the raw drinking water has proven to be of excellent suitability for human consumption and industrial usage. Reasons are the practical absence of xenobiotics, the low nitrate content, the low hardness, the calcite-carbondioxide equilibrium and the constant and low temperature (Stabel, 1998).

The ability to accurately reproduce the physical attributes, the transport and the mixing in a lake, is a prerequisite for water-quality and ecosystem models to simulate biogeochemical processes. A validated ecosystem model could be a valuable tool to assess future development of the Lake Constance ecosystem. Of concern may be the impact climate change could have on the lake due to higher temperatures, different wind speeds or changes in cloud cover. When future targets for nutrient concentrations or loadings are set based on a compromise between the costs of nutrient reduction within the catchment and the desired environmental quality, then the ecosystem model may be used to explore management strategies to meet these desired standards. As a result objective evaluation enables an optimal allocation of (public) financial resources.

The primary objective of this study was to show, that the one-dimensional hydrodynamic model DYRESM can be used to successfully simulate the temporal course of the vertical temperature and salinity distribution in Lake Constance. A second objective was to investigate whether the ecosystem model DYRESM-CAEDYM is capable of reproducing the eutrophication of the lake in the 60s and 70s as well as the recent transition back to an oligotrophic state. Unfortunately the second step could not be addressed within this thesis and remains an item of future research.
CHAPTER 1. INTRODUCTION

Both DYRESM and CAEDYM were developed at the Centre for Water Research (CWR) at the University of Western Australia. They have been applied to many lakes, worldwide, and are updated regularly as a result of continuous research. The models are freely available and can be downloaded from www.cwr.uwa.edu.au/~ttfadmin.

1.2 Lake Constance

1.2.1 Location and Morphometry

Lake Constance (in German: “Bodensee”) is located on the northern edge of the European Alps at a latitude of 47.6°N, where 173 km, 72 km and 28 km shoreline belong Germany, Switzerland and Austria respectively. The altitude of the mean water level is 395.33 m a.s.l. (IGKB, 2002; Wagner et al., 1994). After Lake Geneva, Lake Constance is Europe’s second largest lake in volume (48.5 km³) and can be considered a deep lake with a maximum depth of 253 m. Table 1.1 gives an overview of the lake data.

Lake Constance can be unambiguously subdivided into two main parts: the Upper Lake Constance (“Obersee”) and the much smaller Lower Lake Constance (“Untersee”), which are connected by the 4 km long outflow of the Upper Lake Constance, the Seerhein. The Upper Lake Constance (see also fig. 1.1) has at its north-western end a 18 km long and 2–4 km wide sub-basin, which is called Lake Überlingen (“Überlinger See”). The partitioning of the Upper Lake Constance into Lake Überlingen and the main basin is less distinct, but the sill of Mainau (“Mainau-Schwelle”), a moraine at a depth of 100 m, is a commonly accepted borderline. The main basin has a maximum length of approximately 45 km and a maximum width of 14 km.

The terms Upper Lake Constance and main basin are not used consistently in the literature as some authors use the term “Obersee” if they refer to the main basin only. Furthermore the term “Bodensee” is quite often used (for convenience), if the Upper Lake Constance is meant. This thesis will focus entirely on the Upper Lake Constance.
The basin of Lake Constance was carved by the Rhine Glacier in the last ice-age and water was impounded behind the moraines after the glacier retreated. The Bodanrück, a chain of hills between Lower Lake Constance and Lake Uberlingen, and the sill of Mainau are examples of these moraines.

### Meteorology

Lake Constance lies in the temperate climate zone of central Europe. Mean daily temperatures range from 0.3 °C in January to 18.5 °C in July and the annual mean temperature is 9.2 °C (see section 2.2.3). Peak hourly short wave radiation can reach as high as 950 W/m² on a clear summer day (section 2.2.1).

Mean annual precipitation shows a tremendous increase from 800 to 1400 mm/yr along a longitudinal axis running from the west end to the east end of the lake (Bäuerle et al., 1998). This is because most winds that bring rain are west winds and the west end of the lake is on the lee side of Black Forest, the east end, however, is already on the luv side of the Alps. Wagner et al. (1994) estimate 1002 mm/yr as the mean precipitation on the lake surface.

The large-scale wind field is dominated by winds coming from the south-west, which is orthogonal to the Upper Lake’s longitudinal axis. Local topography can change wind speed and direction. This phenomena is most pronounced in Lake Uberlingen, where the hills (height difference to the lake: > 200 m) along the northeastern and the southwestern shore form a channel. Bäuerle et al. (1998) analysed a 5 year period of wind data and found that, for winds with velocities greater than 3 m/s, there is a general deflection of 20–40° to the right side (clockwise) of winds measured at Friedrichshafen compared to winds measured at Konstanz.

The wind speeds at Konstanz and Güttingen appear to be very similar, while Friedrichshafen generally experiences higher wind speeds. Figure 1.2 shows, for example, that in Friedrichshafen a mean hourly wind speed of 4 m/s is exceeded 25 % of the time, but in Konstanz only 10 % of the time. The mean wind speeds are 1.9 m/s in Konstanz and 3.0 m/s in Friedrichshafen (see also section 2.2.5).

Superimposed on the large-scale wind system are thermally induced, and generally weaker local (land-lake breeze) and regional winds (mountain-valley wind system). The scale of the land-lake breeze is of the same order as the basin and is usually directed perpendicular to the shore. The mountain-valley wind system plays a role only in the eastern part of the basin. Both systems have a diurnal period.

Also restricted to the eastern part of the lake is a warm, dry wind called “Föhn”.

---

Table 1.1: Volume, surface area, mean and maximum depth of the different parts of Lake Constance according to:
a:Heinz (1990), b:Hollan et al. (1990), c:Bäuerle et al. (1998),

<table>
<thead>
<tr>
<th>part</th>
<th>volume (km³)</th>
<th>area (km²)</th>
<th>mean depth (m)</th>
<th>max. depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lower LC</td>
<td>0.8³, 0.9³</td>
<td>71³</td>
<td>13³</td>
<td>40³, 46³</td>
</tr>
<tr>
<td>main basin</td>
<td>44.4³</td>
<td>409³</td>
<td>109³</td>
<td>252³, 253³, 254³</td>
</tr>
<tr>
<td>Lake Uberlingen</td>
<td>3.2³</td>
<td>37³</td>
<td>81³</td>
<td>145³, 147³, 150³</td>
</tr>
<tr>
<td>Upper LC</td>
<td>47.6³, 47.7³</td>
<td>446³, 500³</td>
<td>101³, 107³</td>
<td>252³, 253³, 254³</td>
</tr>
<tr>
<td>LC</td>
<td>48.4³, 48.5³</td>
<td>534³, 538.5³, 571³</td>
<td>85–91³</td>
<td>252³, 253³, 254³</td>
</tr>
</tbody>
</table>
(German for hair-dryer). It is a consequence of a low pressure system north of the Alps and a high pressure system on the south side. The forced air flow loses most of its humidity during the rise on the southern side and thus becomes very dry during the descent on the northern side. As the air mass is usually warm, it very often overflows the cold air sitting on the lake and thus does not affect the lake at all (Bäuerle et al., 1998).

1.2.3 Hydrology

Wagner et al. (1994) tried to estimate the water volumes of the components of the water balance equation for Upper Lake Constance, which is illustrated in the bar diagram of fig. 1.3. Three points to note are that groundwater flows are not mentioned, that rivers clearly dominate the water balance and that the total inflow (347.0 m$^3$/s) almost equals the outflow (347.8 m$^3$/s) as the sum of the other three terms, precipitation (14.2 m$^3$/s), evaporation (9.0 m$^3$/s) and withdrawals (4.4 m$^3$/s), is about zero. The hydraulic retention time, i.e. the lake volume divided by mean discharge of the inflows, is 4.2 years.

The discharge of the Seerhein, the natural outflow on the west end of the lake, is highly correlated with the water level (sec. 2.7). The outflow changes only slowly with time, because the large retention volume of the lake dampens the inflow dynamics.

The size of the catchment area of Lake Constance is 11500 km$^2$ and most of it lies in the Alps. The two largest rivers, Alpine Rhine and Bregenzer Ach, which both enter the lake at the southeastern end, account for 81% of the total inflow (fig. 1.3). As their catchment is alpine their discharge varies strongly with the season, where the period of high discharge is during snow melt from April to July and lowest discharge is from December to February (see also fig. 2.16 and 2.18). A further consequence of their alpine catchment is the quick reaction of the rivers to rain events, so that discharges are highly variable over short time scales. Peak discharge of Alpine Rhine can be as high as 2000 m$^3$/s and minimum discharge as low as 47 m$^3$/s. Flood events are usually accompanied by enormous suspended sediment loads and thus increased density of the river water (see fig. 1.4).

The water levels of Upper Lake Constance display the same seasonal pattern as the
Figure 1.3: The bar chart shows mean discharges of the components of the water balance for Upper Lake Constance (Wagner et al., 1994). The percentages are relative to $Q_{\text{inf}} + Q_{\text{precip}}$. The pie chart illustrates the discharge distribution of the inflowing rivers. Clockwise from the largest fraction: Alpine Rhine, Bregenzer Ach, Argen, Rheintal-Binnenkanal, Schussen, others (own calculations, see also section 2.3).

Figure 1.4: Distribution functions (left panel) and time series (right panel) of suspended solids concentrations in Alpine Rhine. Data from Federal Office for Water and Geology FOWG for stations Schmitter and Diepoldsau.
inflows, except for a small time lag. The monthly maximum is in July and the monthly minimum in February, where the mean difference between these two months is 1.34 m. An analysis of long-term water level records from 1887–1987 showed that, during this time, mean water level decreased by 16 cm and high water levels dropped by 26 cm (Bäuerle et al., 1998).

1.2.4 Hydrodynamics

1.2.4.1 Influence of Salinity on the Stratification

Density stratification in Lake Constance is clearly dominated by temperature differences and to a much lesser extent by variations in conductivity for most months. Temperature differences may be as large as 20 °C with an approximately constant hypolimnion temperature of 4.5 °C. Such a temperature difference would result in a density difference of 2.7 kg/m$^3$.

According to eq. 2.10 an increase in conductivity of 100 µS/cm causes an increase of density of 0.067 kg/m$^3$. The lake water has a mean conductivity of 288 µS/cm, with a minimum of 214 µS/cm and a maximum of 332 µS/cm (calculated from data described in section 2.6.1). Conductivity in the hypolimnion is larger than in the epilimnion by about 50 µS/cm in summer and 10 µS/cm in winter, which results in density differences of 0.034 kg/m$^3$ and 0.007 kg/m$^3$ in the summer and winter respectively. Such differences are insignificant when compared to the density difference that results from the temperature variations mentioned above.

Ollinger (1999) shows however, that at times when the temperatures are between 2 and 6 °C, i.e. the temperature range of small density changes with temperature ($d\rho/d\theta \approx 0$), conductivity differences should not be neglected and therefore must be part of any modelling attempt.

1.2.4.2 Role of Inflowing Rivers

Besides the geometry, the initial density difference and the flow rate govern the inflow dynamics, where it is of prime interest — especially for questions regarding water quality — at which depth the river water intrudes. The conductivity of river water ranges from 200–600 µS/cm, where the large rivers from the alpine regions have lower conductivities (200-350 µS/cm) than the rivers entering on the northern shore (see section 2.3). Therefore, with this small difference in conductivity to the lake water, the initial density difference is mainly due to temperature differences and to suspended solids concentrations. The latter can increase density by several kilograms per cubic metre (fig. 1.4). Suspended solids, however, start to settle out while the river plunges into the lake which is not the case with dissolved ions. This makes the prediction of the insertion depth even more difficult. Observations during a flood event in 1991 gave no hint that river water of Alpine Rhine plunged deeper than 60 m.

Hollan (1998) tried to answer the question, whether Alpine Rhine can drive large surface vortices that are observed in the eastern end of the lake. His calculation suggests that moderate discharges of 500 m$^3$/s can maintain large eddies, but for the largest possible eddy (radius=6 km) discharges at least three times as large are required.

Another influence on the hydrodynamics is that the river water usually has different light attenuation properties and nutrient concentrations. The former may lead directly, the latter indirectly via plankton growth, to different heating in the vicinity of the river mouth and thereby to horizontal temperature gradients.

Zenger et al. (1990) propose that higher conductivity values in Lake Überlingen compared to the main basin are a consequence of the diluting effect of Alpine Rhine flowing
Figure 1.5: Heat im- and exported by rivers, the outflow Seerhein and the drinking water withdrawals, total thermal energy is thermal energy by rivers minus outflow minus drinking water. Lake temperatures are interpolated from data of Internationale Gewässerschutzkommission für den Bodensee. The thermal energy values are only valid for the total flux. See text.

“through” the main basin. This also sets up horizontal gradients, even though they are small.

The thermal energy balance of the system comprising the inflowing rivers, the outflow and the withdrawals is slightly negative (heat is lost). This might be an anticipated result as the temperatures of the inflows are generally a bit colder than the lake surface temperature, but at the same time the majority of the water that leaves the lake, comes from the surface layer. As an example these energy fluxes are plotted in fig. 1.5 for the years 1996 and 1997. Strictly speaking, the absolute values of the heat fluxes of rivers, outflow and drinking water are invalid, as they are dependent on the chosen reference temperature of 0 °C. Nevertheless, they were included to show their dynamics and because relative comparisons are valid. The total flux (or net flux) is the thermal energy import by rivers minus the export by the outflow and the withdrawn drinking water.

An analysis for the period 1980–2000 showed that, for 90 % of the time, the net heat flux is between −56 and +11 W/m² with only a few days per year — corresponding to high discharge events — beyond these limits. If only the mean flux of −14 W/m² is considered, these findings agree with the statement by Bäuerle et al. (1998) that the net heat flux by in- and outflows is negligible compared to the net atmospheric heat flux, where the monthly net atmospheric heat flux ranges from −100 to +110 W/m² (Ollinger, 1999). But as it can also be seen from fig. 1.5, this might not be true for the few days of flood events. The minimum and maximum net heat fluxes are −94 W/m² and +167 W/m² in the period 1980–2000.

Lastly, river inflows are of major importance when it comes to nutrient supply. This can be shown by a simple thought experiment, where it is assumed that the lake is stratified from May to November (6 months) and that no river plunges deeper than 20 m. With the lake volume of the top 20 m (8.3 km³) and the mean discharge of 461 m³/s during that period, the hydraulic residence time of this compartment becomes 6.9 months. In other words, this simplified scenario predicts, that 88 % of the water in the top 20 m will be replaced by new river water within these 6 months, if weak vertical exchange due to strong stratification is assumed.
1.2.4.3 Annual Cycle of Stratification

In most years the water column in Upper Lake Constance attains an almost constant temperature near 4.0–4.5 °C in February or March. An inverse temperature profile is seldom, the formation of an ice-cover is exceptional. In the 21 years from 1980–2000 temperatures at the surface dropped slightly below 4 °C only in the winters of 1985–1987 (see fig. 1.6). The reoccurrence interval of a complete ice-coverage (in German: “Seegfrörne”) might be in the order of 100 years and the last one dates back to February 1963.

Starting in late March or April the heat input generally becomes large enough to build up a weak stratification provided there is a calm period of several days. Bäuerle et al. (1998) illustrate, with an example from March 1990, that the development of the stratification process can differ substantially between Lake Überlingen and the main basin, where a wind event was able to mix the heat down to 70 m in the less sheltered main basin compared to only 20 m in Lake Überlingen. A thermocline was then still present in Lake Überlingen, but had vanished in the main basin. Horizontal gradients are the consequence of this process.

In May or June, during the time of highest net heat fluxes (fig. 2.12 in Ollinger (1999) and fig. 2.31 on p. 60), stratification becomes strong enough to resist the complete erosion of the thermocline. Strong wind events will now cause the thermocline to tilt and upwelling at the upwind end may occur. See for example fig. 14 in Bäuerle et al. (1998), where upwelling of the thermocline from a depth of 7 m occurred in Lake Überlingen on
1.2. LAKE CONSTANCE

9 May. This coincides with the minimum of the modified Lake Number $L_N$ plotted in fig. 1.7, which drops to almost unity. The calculation of the modified version of the Lake Number is explained in sec. 1.3.2.1 on p. 11.

Between the end of July and the end of August, as the net flux approaches zero, the surface temperature maximum is reached, which ranges from 19 °C in 1989 to 24 °C in 1992, 1994 and 1995 (fig. 1.6). Apart from a few exceptions, the modified Lake Number is much greater than one ($\approx 10^{-70}$), indicating that the thermocline remains mainly horizontal and internal wave activity is relatively low as the water column stability is too high to cause agitation.

From September to late autumn the thermocline deepens and upwelling events will happen. The deepening is a result of the increased turbulence in the surface layer which is caused by the combined action of more vigorous winds in autumn and strong convective overturn as the net heat flux reaches its yearly minimum. The intense cooling reduces the potential energy of the lake, so that even minor storms can cause upwelling.

Two beautiful examples are depicted in figs. 16 and 17 in Bauerle et al. (1998). Two stronger wind events ($U \approx 8 \text{ m/s}$) within a couple of days caused internal seiching ($T \approx 3 \text{ days}$) to increase in amplitude from 4 to 20 m on 2 September 1992 and upwelling occurs on 5 September. The other example on 27 October, two month later, showed an upwelling of water from 30 m depth when stability was already weak. These two examples correspond to modified Lake Numbers of 3 and 0.5 respectively in fig. 1.7. While turbulence by convective overturn will erode the thermocline in a continuous process, a single strong storm event may now be able to cause complete overturn of the lake in one step.

1.2.4.4 Enhanced Mixing Near the Sill of Mainau

The sill of Mainau divides the main basin from Lake Überlingen and is a location of minimum cross-sectional area. In many situations, the vertical structure of currents above the sill is a two layer flow divided by the thermocline and with opposite flow directions. The currents are forced by wind, which drags water in the surface layer downwind and induces a current in upwind direction in the bottom layer. It is expected that the constriction of the flow area causes higher velocities and the increased shear may enhance vertical mixing, if the Gradient Richardson Number $R_i$ becomes supercritical ($R_i \leq 1/4$).

Measurements by Boehrer et al. (2000) at the sill of Mainau revealed that $R_i$ is supercritical for short periods of time. From their measurements during mainly southwesterly to northwesterly winds, they concluded that vertical transport at the sill may surmount the transport of the entire remaining lake, even though the region at the sill accounts for only 1% of the lake area. This is similar to the result found by Heinz (1990) and Heinz et al. (1990) with the flux gradient method. Kocsis et al. (1998) calculated that vertical transport above the sill should be 34 times larger than the basin wide average, which is derived from measurements during northeasterly winds.

The analysis of data collected from eight thermistor chains moored in Upper Lake Constance from 11 October to 17 November 2001 (Appt, 2002), confirmed earlier observations that intrusions of water of intermediate density occur in Lake Überlingen and propagate into the main basin at a depth of 20-50 m.
1.3 The Hydrodynamical Model DYRESM

1.3.1 Overview

The acronym DYRESM stands for Dynamic Reservoir Simulation Model. It is a one-dimensional numerical model to simulate the vertical temperature, salinity and density structure in lakes and reservoirs. It can be coupled to the aquatic ecosystem model CAEDYM, which is also available for free at the Centre for Water Research. Development of DYRESM dates back to 1978 with a first successful application on Wellington Reservoir in Western Australia (Imberger et al., 1978; Imberger & Patterson, 1981). It has since been applied — partly with modified codes — to many other reservoirs and small lakes, e. g. Prospect Reservoir in Australia (Schladow & Hamilton, 1997), Kootenay Lake in Canada (Patterson et al., 1984). Examples for simulations on large lakes are Ivey & Patterson (1984) on Lake Erie and Hollan et al. (1990) for a period of 14 months on Upper Lake Constance. DYRESM can be considered as a validated model.

As input data, DYRESM requires the lake geometry, daily or subdaily meteorological data, where subdaily resolution must be between 15 min and 3 hr, daily discharge and temperature of inflows, daily discharge of outflows or withdrawals and finally the light extinction coefficient averaged in depth and time. The data requirements are described in

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Possibly unique to DYRESM, compared to the hydrodynamical components of other lake models such as MINLAKE or CE-QUAL, is that virtually no calibration is required. This means not only that all relevant physical processes occurring in a lake are included, but also their level of description is fundamentally correct. The parameterisation is based on detailed field and laboratory studies (Hamilton & Schladow, 1997).

Another important feature of DYRESM is the layer concept, in which the reservoir is modelled as a system of dynamic, horizontal layers of uniform property, as opposed to a fixed grid approach in which differential equations are solved numerically on the mesh points. These layers can move up or down due to inserted inflows or withdrawals below, thereby changing their thickness according to the volume-height relationship. Mixing and surface layer deepening is modelled by amalgamation of layers based on a criterion of available turbulent kinetic energy and required potential energy. Diffusion in the meta- and hypolimnion is simulated by transferring volume fractions from one layer to the overlying layer, where the fraction is determined by the Lake Number (not yet implemented in the official release). The layer concept is computationally relatively simple, thus long-term simulations become feasible, and it greatly reduces numerical diffusion and stability problems that are associated with a fixed grid technique (Imberger & Patterson, 1981; Hamilton & Schladow, 1997).

1.3.2 Assumption of One-Dimensionality and Lake Constance

The fundamental assumption of an one-dimensional lake model is that horizontal gradients are far smaller than vertical gradients, or in other words, properties like temperature and salinity are laterally homogenous. This assumption is valid if a density stratification is present, disturbing forces originating from wind, in- and outflows are weak and other processes that also generate horizontal gradients are negligible. Stratification, which lasts in most lakes for about 9 months, inhibits vertical motions and reduces turbulence while horizontal gradients are relaxed on time scales of less than a day (Imberger & Patterson, 1981; Antenucci & Imerito, 2001).

The influence of wind and in- and outflows can be estimated from non-dimensional numbers. The influence of the wind can be described by the Wedderburn Number and the Lake Number (Imberger & Patterson, 1990). Internal Froude Numbers have been used to judge the effect of in- and outflows on the vertical structure (Patterson et al., 1984).

1.3.2.1 Lake Number Criterion

In the simplified case of a rectangular basin and two layers, the Wedderburn Number is equal to the Lake Number (Imberger, 1998). Thus, for simplicity, only the Lake Number \( L_N \) will be considered in the following. From the field data of temperature and conductivity (sec. 2.6.1 on p. 54) and the lake geometry (sec. 2.1 on p. 18), a quantity \( S \), that can be interpreted as a stratification strength or as the Lake Number at unit wind stress, was calculated

\[
S = \frac{(z_V - z_G) \cdot G \cdot \beta_{tc}}{A \cdot d_{CoV}} = L_N \cdot \tau
\]

where the term \((z_V - z_G)\) is the vertical distance between the centre of volume and the centre of gravity, \(G\) is the gravity force exerted on the lake, \(A\) is the surface area on which the mean daily wind stress \(\tau\) acts upon and \(d_{CoV}\) is the depth to the centre of volume. In the definition of the Lake Number by Imberger & Patterson (1990), \(\beta_{tc}\) is the slope of the
thermocline (strictly speaking the pycnocline), which would cause upwelling of water from the hypolimnion. However, deviating from that definition the slope \( \beta_{tc} \) was fixed to value of 0.001 for two reasons: the unambiguous determination of the thermocline depth in a computer algorithm is a very ambitious and difficult task and secondly the field data have a relatively coarse spatial resolution. The value of 0.001 was chosen as this corresponds to a 20 m vertical displacement of the thermocline on a horizontal distance of 40 km. During the stratified season, 20 m is the approximate depth of the lower metalimnion and 40 km is about the length scale of the main basin.

The quantity \( S \) contains only constants and stratification properties, and if it is assumed that density stratification changes smoothly between field data days, then \( S \) can be confidently interpolated between field data days. The interpolated daily values of \( S \) are then divided by the mean daily wind stress \( \tau \), which is calculated from the prepared wind data (sec. 2.2.5 on p. 32). This results in a Lake Number that is here referred to as the “modified Lake Number” \( L_N^\beta \) due to the fixed thermocline slope.

The modified Lake Number is plotted versus time in fig. 1.7. Large values \( (L_N^\beta \gg 1) \) indicate that restoring forces by the stratification are stronger than overturning forces by the wind and vice versa \( (L_N^\beta < 1) \). It can be seen, that \( L_N^\beta \) hardly falls below 2 in the period 01 April to 01 October, indicating that severe deviations from the assumption of a horizontal layer structure are exceptions. In the last quarter of the year, when stratification has weakened and storms are likely to occur, \( L_N^\beta \) frequently becomes smaller than unity. An interpretation of a value of unity is that an equilibrium between the restoring momentum by the stratification and the overturning momentum by the wind stress is achieved at a pycnocline slope of 0.001.

In summary, it can be expected that model algorithms can successfully simulate the stratification until the first severe storms in autumn cause strong upwelling and mixing within a short time, which might not be resolved by the model.

1.3.2.2 Internal Froude Number Criterion

The internal Froude Number \( F \) is the ratio of the inertial forces associated with the inflows or withdrawals and the pressure forces induced by horizontal density gradient generated by the inflow or withdrawals. For \( F < 1 \) the assumption of one-dimensionality is justified.

The internal Froude Numbers \( F_{inf} \) and \( F_{wdr} \) for inflows and withdrawals respectively, are defined as (Patterson et al., 1984)

\[
F_{inf} = \frac{u}{(g' \cdot d)^{1/2}}
\]

\[
F_{wdr} = \frac{Q}{d^2 \cdot (g' \cdot d)^{1/2}}
\]

where \( u \) is the inflow velocity, \( g' = \Delta \rho / \rho \cdot g \) is the effective acceleration of gravity, \( \Delta \rho \) is the density difference between river and lake surface or withdrawal level and surface respectively, \( Q \) is the withdrawal rate and \( d \) is the reservoir depth.

With an inflow velocity of \( u = 1 \) m/s (Alpine Rhine at a discharge of 500 m\(^3\)/s from Hollan (1998)), mean lake depths \( d \) of 100 m for \( F_{inf} \) (main basin) and 80 m for \( F_{wdr} \) (Lake Überlingen ), \( g' = 9.81 \cdot 10^{-3} \) m/s\(^2\) and \( Q = 4.0 \) m\(^3\)/s, \( F_{inf} \) becomes 1.0 and \( F_{wdr} \) becomes 7.1 \cdot 10^{-4}.

Thus, the withdrawal in Lake Überlingen appears to have a negligible effect on the stratification, whereas the value of \( F_{inf} \) confirms that the inflow of Alpine Rhine leads to three-dimensional effects (e. g. fig. 10 in Bäuerle et al. (1998)), which will not be captured by DYRESM.
1.3.2.3 Other Processes

Other processes that are beyond the scope of a one-dimensional model are differential heating, i.e. shallow regions along the shore heat up or cool down quicker than the interior, and wind sheltering. Both lead to horizontal gradients, which increase mixing, and are more important in small than in large lakes. Zenger et al. (1990) conclude that wind sheltering and consequently less vertical mixing in Lake Überlingen during winter is, among other reasons, responsible for building up a horizontal gradient towards the main basin.

1.3.3 Outline of Main Subroutines

Of the many subroutines in DYRESM, five of them deal with the key processes surface heat fluxes, mixed layer deepening, turbulent diffusion, inflow insertion and withdrawals. The first two will be explained below, the turbulent diffusion algorithm is currently under development and is described in sec. 2.8.1 on p. 57. For the inflow and withdrawal routines, the reader is referred to Antenucci & Imerito (2001).

Other subroutines can be considered as service routines, e.g. for reading input data, writing output, calculating density from temperature and salinity, computing layer thickness from layer volume or area and vice versa.

Surface heating and mixed layer deepening are both calculated with a sub-daily time step whereas the algorithms for inflows, outflows and turbulent diffusion are only executed once a day. If the meteorological data are sub-daily, then the heating and mixing will have the same time step as the meteorological data. Otherwise the time step of these two subroutines is given by criteria that limit the shear velocity and the maximum temperature change of the uppermost layer.

1.3.3.1 Surface Heat Fluxes

The following heat fluxes are considered: sensible heat flux, latent heat flux due to evaporation or condensation and the heat fluxes by short and long wave radiation. Heat exchange with the sediment is currently not implemented, but is being developed by Katherine Prescott from CWR.

**Sensible and Latent Heat Flux** The sensible and latent heat fluxes, $q_s$ and $q_L$, are described by bulk aerodynamic formulae (Antenucci & Imerito, 2001)

\[
q_s = C_S \cdot \rho_a \cdot C_p \cdot U \cdot (\theta_a - \theta_w) \tag{1.4}
\]

\[
q_L = 0.622 \cdot P \cdot C_L \cdot \rho_a \cdot C_{eva} \cdot U \cdot (e_a - e_s(\theta_w)) \tag{1.5}
\]

where $C_S$ and $C_L$ are transfer coefficients (see sec. 2.5.1.1), $\rho_a$ is the air density, $U$ is the wind speed, $\theta_a$ and $\theta_w$ are the air and surface water temperatures, $C_P$ is the heat capacity of water at constant pressure, $P$ is the atmospheric pressure [mbar], $C_{eva}$ is the latent heat of evaporation and $e_a$ and $e_s$ are the air vapour pressure and the saturation vapour pressure [mbar], which is a function of the surface water temperature. An additional restriction is that heat gain by condensation is not allowed. Energy transferred according to these equations is always added or subtracted to the uppermost layer.

**Short Wave Radiation Flux** The incoming solar short wave radiation flux $q_{sw}^{in}$ is partitioned into two fractions, a penetrative wavelength range from 280–700 nm and a non-penetrative range from 700–2800 nm. About 55 % of the energy is associated with the latter fraction and is absorbed (after partial reflection) in the uppermost layer. The
remaining 45% of the radiation energy is allowed to penetrate the water column and is absorbed according to Lambert-Beer’s law:

\[ q_{sw, <700}(z) = 0.45 \cdot (1 - r_a^{(sw)}) \cdot q_{sw}^{in} \cdot e^{-k_{lw} \cdot z} \]  

(1.6)

where \( q_{sw}^{in} \) is the incident short wave radiation flux [e. g. W/m\(^2\)], \( r_a^{(sw)} \) is the albedo (sec. 2.5.1.2), \( k_{lw} \) is the depth averaged light extinction coefficient (sec. 2.5.3) and \( z \) is the space coordinate (positive in downward direction).

If the meteorological data are of daily resolution only, the short wave radiation flux is assumed to have a sinusoidal distribution over the day (eq. 2.3 on p. 25).

**Long Wave Radiation Flux** For the case that direct long wave measurements are not available, the long wave radiation flux \( q_{lw} \) emitted from the lake is calculated in accordance to the Stefan-Boltzmann law for black body radiation:

\[ q_{lw, out} = \varepsilon_w \cdot \sigma \cdot \theta_w^4 \]  

(1.7)

where \( \varepsilon_w \) is the emissivity of the water (sec. 2.5.1.3), \( \sigma \) is a physical constant \( (5.6696 \cdot 10^{-8} \text{ W m}^2 \text{ K}^{-4}) \) and \( \theta_w \) is the absolute temperature [K] of the water surface. This heat loss affects the uppermost layer only.

The incident long wave radiation flux, which is absorbed completely in the uppermost layer, can be estimated from cloud cover fraction CCF and air temperature:

\[ q_{lw, in} = (1 - r_{lw}) \cdot (1 + 0.17 \cdot CCF^2) \cdot \varepsilon_a \sigma \theta_a^4 \]  

(1.8)

\[ \varepsilon_a = C_\varepsilon \cdot \theta_a^2 \]  

(1.9)

where the subscripts a and w denote air and water respectively, \( r_{lw} \) is a reflection coefficient (assumed to be a constant value of 0.03), \( \varepsilon_a \) is the emissivity of the atmosphere, and \( C_\varepsilon \) is a constant of proportionality with a value of \( 9.37 \cdot 10^{-6} \text{ K}^{-2} \) (Antenucci & Imerito, 2001).

**1.3.3.2 Surface Layer Deepening** After the heating routine, the mixed layer deepening routine is invoked. It is assumed that turbulent kinetic energy (TKE) is generated by the three mechanisms (1) convective overturn, (2) wind stirring and (3) shear flow between layers. At each time step, mixing by these three mechanisms is modelled separately in the above order (Antenucci & Imerito, 2001; Imberger & Patterson, 1981).

1. **Mixing by Convective Overturn**

Beginning from the uppermost layer, the layers are checked for instabilities which may result from surface cooling. If an instability between two layers exists, they will be merged and a fraction of the potential energy released will become available as \( TKE_{PE} \). The efficiency of this conversion is set by the value of the Potential Energy Mixing Efficiency \( \eta_{PE} \) (sec. 2.5.1.7 on p. 51).

Furthermore, whenever two layers are merged, of which at least one has a non-zero mean layer speed due to the acceleration by wind stress at the previous time step(s), energy is released from the kinetic energy (KE) inherent in the mean horizontal velocity of the layers. This can be shown by calculating the change in mechanical energy under conservation of momentum (e. g. Antenucci & Imerito (2001)). The fraction \( \eta_{KE} \) of the released KE becomes available as \( TKE_{KE} \), where \( \eta_{KE} \) is termed Shear Production Efficiency (sec. 2.5.1.6).
2. Mixing by Wind Stirring

Wind stirring during the time step $\Delta t$ leads to the production of $TKE$ in the uppermost layer

$$TKE_{ws} = \eta_{ws} \cdot \rho_w \cdot u_*^3 \cdot \Delta t \quad (1.10)$$

where $\eta_{ws}$ is the Wind Stirring Efficiency (sec. 2.5.1.8), $\rho_w$ is the density of the water in the uppermost layer, $u_*$ is the friction velocity and $\tau$ the wind stress. The wind stress is calculated from an aerodynamic bulk formulae from the wind speed $U$

$$\tau = C_M \cdot \rho_a \cdot U^2 \quad (1.11)$$

where $C_M$ is the transfer coefficient for momentum (sec. 2.5.1.1) and $\rho_a$ the air density. $TKE_{ws}$ from this time step, plus any $TKE$ left from the previous time step, plus what was possibly generated by convective overturn $TKE_{PE} + TKE_{KE}$ forms the $TKE_{av}$ available for further mixing. Starting from the uppermost layer, the energy required to mix a layer with the layer underneath, $E_{req}$, is compared with $TKE_{av}$. The required energy is, as a first approximation, the increase in potential energy. However, as already explained at point (1), it will be less due to the partial conversion of KE into TKE, if at least one of the layers possesses KE from a mean flow. If $TKE_{av} > E_{req}$, the two layers will mix, the remaining $TKE_{av}$ will be computed and the next pair of layers will be compared until the available $TKE$ becomes too small for any further deepening. Otherwise the code proceeds directly to point (3).

3. Mixing by Shear Flow

At this point, the two mechanisms above have usually already deepened the mixed surface layer so that this last mechanism can be best interpreted as the deepening by shear at the base of the mixed layer. This is done by calculating the mean horizontal velocity $u$, that the uppermost layer will have attained at the end of the time step $\Delta t$ due to the acceleration by wind

$$u = u_*^2 / h \cdot \Delta t + u_{old} \quad (1.12)$$

where $h$ is the thickness of the layer and $u_{old}$ is the mean layer velocity from the previous time step after possible adjustments due to the mixing in (1) and (2). However, acceleration only commences when the wind speed $U$ exceeds a critical wind speed $U_{crit}$. Additionally, the duration of acceleration is limited by the shear period $T_{sp}$. Both the critical wind speed and the shear period are explained in sec. 2.5.2.2.

Then analogous to above, the required energy is compared to available $TKE$, where the comparison starts again at the uppermost two layers. Of course, $TKE_{av}$ is the remaining $TKE_{av}$ of the deepening by wind stirring (2). But, $E_{req}$ is less than before as the $KE$ of the wind accelerated layer has increased. Deepening will stop if $TKE_{av} < E_{req}$. 

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Chapter 2

Methods

This chapter describes what kind of data are required to model the stratification of Lake Constance with DYRESM, what assumptions were made when preparing the final input data files from the raw data and what kind of modifications to the code were made in order to enhance the model results. The sections about the data preparation are intentionally very detailed. The main reason is that future studies of Lake Constance with DYRESM-CAEDYM can only rely on already prepared data\footnote{save that all institutions that provided data will allow further usage of the data within research projects}, if it is exactly known from where the data came, if all available data were collected and what the assumptions and modifications were. It may also serve as a guide in the preparation of additional data.

As with any numerical simulation, four types of data are required: (1) geometry of the system, (2) forcing variables, (3) parameters and (4) the state variables, where the first three can be referred to as model input data.

*Geometric* information comprises the 1D-bathymetry, details of the stream bed downstream of a river mouth and elevations of outflows, withdrawals and spillways (sec. 2.1).

The *forcing* variables “drive” the system, as they determine the energy, momentum and mass fluxes across the system boundary. Forcing variables in DYRESM are the meteorological variables (section 2.2), discharge, salinity and temperature of inflowing rivers (section 2.3), and discharge of outflowing rivers or withdrawals (section 2.4). It was attempted to collect meteorological data in a hourly resolution and all other forcing data in a daily resolution for the 41 year period from 1960–2000.

*Parameters* appear in empirical equations. They are usually determined by field measurements or lab experiments or are subject to a calibration process. DYRESM has only a few parameters that can be specified by the user. Most of them should be considered as generic constants as they appear to be unspecific to a certain location or geometry. The only intrinsic calibration parameters are the diffusion volume fraction which is used to simulate internal mixing and the benthic boundary layer thickness which is used to model benthic boundary mixing. The light extinction coefficient might need calibration if field measurements are not available or very uncertain. The parameter settings will be discussed in section 2.5.

*The state variables* are the variables that describe the state of a system. They are the unknowns for which the numerical model solves. In DYRESM these are temperature and salinity (section 2.6) as well as the water level (section 2.7). These data are required to initialise the model and to verify the model computations. Verification is inevitable to
A few changes were made to the current web-release version of DYRESM-CAEDYM code (version 2.2.0-beta2, released 20 Dec. 2001). The first change was to allow for negative air temperatures in the input data. Secondly, modelling a deep lake gave the incentive to develop an algorithm which would permit thick layers in the deeper parts of the lake, but enforce thin layers in the near surface region. Lastly, model results of the web-release version indicated that DYRESM underestimates mixing in the metalimnion. This led to the inclusion of algorithms that execute internal and boundary layer mixing. Section 2.8 explains these changes in detail.

### 2.1 Geometric Information

DYRESM requires an input file ("Storage File") which provides information about the latitude of the lake, the details of the inflows and outflows and the 1-D bathymetry.

The latitude of the main basin of Lake Constance is 47.6°N. If the resolution of the meteorological data is daily, the latitude will be used to generate the short wave radiation distribution during a day.

To determine the insertion depth of inflowing rivers, the model requires for each river:

1. the average slope along the longitudinal axis,
2. half the base angle of the triangular shaped cross-section
3. the stream bed drag coefficient
downstream of the river mouth. The drag coefficient $C_D$ was assumed to be 0.015 throughout (Antenucci, 2001). Table 2.1 shows the slope and half angle as they were estimated from a topographic map with 25 m contour intervals (Landesvermessungsamt Baden-Württemberg, 1991). Groundwater flow into or out of Lake Constance appears to be an insignificant term in the water balance as it was not included in the water-budget calculations by Wagner et al. (1994). Therefore groundwater flows were not considered in this thesis. However, for certain questions regarding water quality, the relevance of groundwater discharge may have to be more closely investigated.

The Seerhein is the lake’s only outflow. A outflow elevation of 393.5 m a.s.l. was chosen, which is approximately 0.5 m less than the minimum observed water level (site inspection, own calculations). For convenience of modelling, the numerous drinking water withdrawals were combined into one withdrawal with a constant flow rate of 5.5 m$^3$/s (Stabel & Kleiner, 1995). The Zweckverband Bodensee-Wasserversorgung operates the largest single withdrawal (4.0 m$^3$/s), which is located in Lake Überlingen at a depth of 60 m (ZVBWV, 2002). Its intake depth was considered as representative for the combined discharge. Lastly an artificial surface outflow was introduced to correct wrong inflow and outflow data. The outflow height is the same height as Seerhein.

Heinz (1990) provides values of volume and area (in the horizontal plane) for 25 layers, each of 10 m thickness, in Lake Constance. It was assumed that the areas given were mean areas at the centre of each layer and that the origin of the depth coordinate was at the mean water level of the lake (IGKB, 2002). Another assumption was that the deepest layer could be enlarged to a thickness of 13 m in order to have a lake depth of 253 m, which is more consistent with the depth of 254 m published by IGKB (2002) and with Bäuerle et al. (1998) who report 253 m.

Using piecewise cubic interpolating polynomials, Heinz’s values were inter- and extrapolated to obtain a 1 m spacing. The chosen interpolation method smoothes data, but does not lead to overshooting or oscillation. Extrapolation was necessary for area values above 390 m a.s.l. and for cumulated volume values above 395 m a.s.l. The areas and volumes are plotted in fig. 2.1.

<table>
<thead>
<tr>
<th>river</th>
<th>half angle $\alpha$ [$^\circ$]</th>
<th>slope $\phi$ [$^\circ$]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alpine Rhine</td>
<td>88</td>
<td>0.95</td>
</tr>
<tr>
<td>Bregenzer Ach</td>
<td>87</td>
<td>0.87</td>
</tr>
<tr>
<td>Argen</td>
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<tr>
<td>Schussen</td>
<td>89</td>
<td>2.66</td>
</tr>
<tr>
<td>Dornbirner Ach</td>
<td>88</td>
<td>0.83</td>
</tr>
<tr>
<td>Leiblach</td>
<td>89</td>
<td>0.72</td>
</tr>
<tr>
<td>Seefelder Aach</td>
<td>89</td>
<td>3.02</td>
</tr>
<tr>
<td>Rotach</td>
<td>89</td>
<td>2.03</td>
</tr>
<tr>
<td>Stockacher Aach</td>
<td>88</td>
<td>0.98</td>
</tr>
<tr>
<td>Goldach</td>
<td>89</td>
<td>1.60</td>
</tr>
<tr>
<td>Steinach</td>
<td>89</td>
<td>2.95</td>
</tr>
<tr>
<td>Aach</td>
<td>89</td>
<td>2.23</td>
</tr>
<tr>
<td>Residual flow</td>
<td>89</td>
<td>2.00</td>
</tr>
</tbody>
</table>

Table 2.1: Geometry of stream bed downstream of river mouth
2.2 Meteorological Data

For the calculation of the heat and mass fluxes at the water surface and for the calculation of the wind shear, DYRESM requires an input file ("Met File") which contains a time series of six meteorological variables:

1. short wave radiation (sec. 2.2.1)
2. long wave radiation or cloud cover fraction (sec. 2.2.2)
3. air temperature (sec. 2.2.3)
4. vapour pressure (sec. 2.2.4)
5. wind speed (sec. 2.2.5)
6. precipitation (sec. 2.2.6)

In the case where direct long wave radiation measurements are not available, cloud cover fraction measurements must be provided instead. The prepared met-file for Lake Constance starts on 01.01.1960 00:00 and ends on 31.12.2000 23:00 with a time step of 1 hour (359424 points in time).

Meteorological data are measured by the three federal weather services of Germany, Switzerland and Austria in direct vicinity to the lake and by the University of Constance with a station in Lake Überlingen (Ollinger, 1999). Other data sources are not known to the author.

Deutscher Wetterdienst (DWD) operates stations in Konstanz (elevation of station: 443 m a.s.l.), Friedrichshafen (401 m a.s.l.) and in Eriskirch-Mariabrunn (408 m a.s.l.). MeteoSwiss (MCH) has a station in Güttingen (440 m a.s.l.) and the Central Institute of Meteorology and Geodynamics (ZAMG) has one in Bregenz. Their locations are shown in fig. 1.1 as black points.

DWD and MCH were asked to provide data for the period from 1960–2000 in hourly resolution except precipitation height for which a daily resolution was deemed sufficient. The obtained data include all that are available in electronic form.

Since hourly data were only available after 1988, data were not requested from ZAMG for station Bregenz, even though they might have helped to get an idea of the spatial variability across the longitudinal axis of the lake.

According to Ollinger (1999), the 'Limnologisches Institut' of University of Constance has records of radiation, temperature, humidity, wind speed and wind direction from a station in the middle of Lake Überlingen since 1986. However, he uses data of station Konstanz because records of the lake station have too many data gaps. Also, as it is uncertain how long the lake station will stay in operation, he argues that using data from station Konstanz ensures that his work is more comparable with future research as consistent data sets can be used. For this thesis University of Constance was not asked to provide data, mainly because records of weather stations on shore are complete and redundant for the period since 1986. Furthermore, it was thought that the station in Lake Überlingen may not represent the climate in the main basin much better than on-shore station as Lake Überlingen differs quite substantially from the main basin (Bäuerle et al., 1998).

If DYRESM is set to simulate conditions under non-neutral atmospheric stability, the height above lake bottom of the meteorological station must be known. The majority of data were from Konstanz, Friedrichshafen and Güttingen, therefore the height was chosen to be 33 m above lake bottom, which is an average of the three stations.
2.2. METEOROLOGICAL DATA

Figure 2.2: Mean daily short wave radiation for each month and the whole year (filled bars). Errorbars indicate the interval in which 90 % of all values fall. Data from Deutscher Wetterdienst (DWD) and MeteoSwiss (MCH) for a time period of 21 years.

2.2.1 Short Wave Radiation

The fraction of the solar radiation that reaches the ground is called incident short wave radiation (SWR) and has a wavelength band of ~280–2800 nm. The dimensions are energy per (horizontal) unit area and time. The remaining part of the solar radiation is absorbed, reflected or back-scattered by the atmosphere.

Short wave radiation was provided in an hourly resolution by DWD and MCH at the stations Konstanz and Guttingen respectively. DWD supplied short wave radiation from 01.01.1980–31.12.2000, MCH from 01.01.1981–31.12.2000. Daily means of SWR for each month and for the whole year are plotted in fig. 2.2.

In short, the following 5 steps were undertaken to create the final hourly time series of 41 years of short wave radiation from the raw data:

1. Treatment of uncertain measurements and erroneous values.
2. Time shift to Central European Time and conversion of units.
3. Interpolation onto time series at full hours.
4. Averaging between stations Konstanz and Guttingen.
5. Regression between cloud cover fraction and SWR in order to estimate the missing 20 years from 1960–1979.

2.2.1.1 Treatment of Uncertain Measurements and Erroneous Values

The Deutscher Wetterdienst flagged an uncertain value by adding 500 J/cm² to the uncertain measurement. Because maximum radiation is always below 500 J/cm², all uncertain measurements are easily recognised as the ones equal to or larger than 500 J/cm². For convenience it was assumed that the uncertain measurements can be used without checking each for their plausibility.

MeteoSwiss had slightly negative radiation values in its time series. These values were set to zero.

\[10/1977–31.12.1979\] is also available but was not delivered, even after repeated inquiries.
A closer look at the largest values in the time series of short wave radiation raised the suspicion that they might be erroneous. Fig. 2.3 shows the cumulative probability of short wave radiation. It seems that values larger than approximately 950 W/m$^2$ might be due to measurement errors and were therefore set to 950 W/m$^2$.

### 2.2.1.2 Time Shift and Conversion of Units

DWD’s measurements were given in units of J/cm$^2$ for the end of a one hour measuring period where the time was given as apparent solar time$^3$ (AST). The conversion from AST to Central European Time (CET) is (priv. comm. DWD, Klima- und Umweltberatung, Hamburg)

$$\text{CET} = \text{AST} + \text{timezone} - \lambda \cdot 1/\Omega - \zeta$$

(2.1)

where timezone is the time zone difference between CET and Greenwich Mean Time (i.e. +1 hr), $\lambda$ is the longitude (i.e. 9.17 deg), $\Omega = 15$ deg/hr and $\zeta$ is the difference between AST and mean solar time. The AST varies by $-14$ to $+16$ min around the (local) mean solar time during a year due to the eccentricity of the earth’s orbit and the inclination of the ecliptic to the celestial equator (Isaacs et al., 1999). $\zeta$ was neglected in the conversion, because a quarter of an hour was not considered as important.

MeteoSwiss’ measurements were given in units of Wh/m$^2$ for the end of a one hour measuring period where the time was given in UTC$^4$. The time was converted to CET by adding 1 hour.

With both DWD and MeteoSwiss, the value of a measurement over an hour period was reported at the end of that period. This will be referred to as “time stamp at the end” in the following text. However, since time stamp at the beginning was used, 1 hour had to be subtracted.
2.2.1.3 Interpolation onto Time Series at Full Hours

Values at Güttlingen were consistently measured 40 min after any full hour, values at Konstanz 23 min after any full hour. But, since the final time series is at every full hour, either a time shift of the complete data or an interpolation was required. Linear interpolation was preferred to a time shift even though it reduces extremes while conserving the energy (see also fig. 2.4). The maximum data gap size that was allowed to be interpolated was two hours, i.e. one missing value.

2.2.1.4 Averaging between stations Konstanz and Güttlingen

The mean daily SWR is 11.0 MJ/m² in Konstanz and 10.8 MJ/m² in Güttlingen, which is a difference of less than 2%. This difference seems to be too small to claim that a systematic bias between the stations exists, rather it was considered random. Thus, for every hour the values of short wave radiation were averaged between the stations. The resulting time series has still two data gaps of four and eight missing hours respectively. These missing values were estimated from cloud cover as described below.

2.2.1.5 Regression between cloud cover fraction and short wave radiation

The missing short wave radiation measurements from 1960–1979 required the following procedure to estimate SWR from cloud cover fraction (CCF):

1. Developing a sinus function to calculate the daily radiation energy on days without clouds for every day of the year.

2. Finding a sinus function that distributes this daily energy over a day.

3. Employing regression analysis to find the relation between CCF and the radiation ratio, which is the ratio between actual radiation and radiation without clouds.

The first two points made it possible to calculate, for every hour of the year, the SWR that would be expected for cloud-free conditions, the last point made it possible to calculate the fraction of the cloud-free conditions.

---

3AST is a local time that would be shown by a sundial that is calibrated such that the sun is highest at noon
4Coordinated Universal Time
Equation 2.2 describes daily short wave radiation energy $E_{sw}$ as a function of the day of the year $JD$ and the parameters amplitude $A$, period $T$, time shift $t_{sh}$ and mean SWR energy $E_{mean}$.

$$E_{sw}(JD) = A \cdot \sin\left(\frac{2\pi}{T}(JD - t_{sh})\right) + E_{mean}$$  \hspace{1cm} (2.2)

The amplitude is, in theory, half the difference between radiation energy on 21st of June and 21st of December, where each of these days would need to be a perfectly cloud-free day. A day was conceived as cloud-free if at least three observations of CCF during daylight hours had a value of zero. Daylight hours were defined as the time between 6am and 9pm. Each of the two stations Konstanz and Göttingen were filtered for these criteria independently. There were hardly any cloud-free days on the dates mentioned above, hence, the criterion was relaxed to the time periods from 12.–30. December and 15.–27. June, which then yielded a sufficient number of days (see fig. 2.5, a and b). The amplitude was then calculated from the maximum radiation values found in each of the two periods ($A = 1.187 \cdot 10^7$ J/m$^2$). The value of $E_{mean}$ is $1.814 \cdot 10^7$ J/m$^2$, which was just the average of the minimum and maximum radiation. The period $T$ followed from the annual cycle ($T = 365.25$ days). The time shift $t_{sh}$ had to be such that peak radiation occurs on 21st of June which is on average day 172.25 of a year (173 on a leap year, 172
2.2. METEOROLOGICAL DATA

Thus \( t_{sh} = 172.25 - T/4 \).

The resulting curve is plotted in the fig. 2.5c together with all occurrences of cloud-free conditions in the period of available short wave radiation data.

**Sinus function of radiation during the day** The next step was to go from daily maximum radiation to hourly maximum radiation, where maximum radiation is assumed to be the one occurring under cloud-free conditions. To do so, another sinus function (eq. 2.3) was required which would distribute a daily radiation energy into mean hourly radiation fluxes \( q_{sw} \) (modified from Antenucci & Imerito, 2001):

\[
q_{sw}(t_1, t_2) = \frac{E_{sw}}{2(t_2 - t_1)} \cdot [\cos(\omega \cdot t_1 - \beta) - \cos(\omega \cdot t_2 - \beta)] \quad (2.3)
\]

\[
\omega = \pi / (\gamma \cdot 86400 \text{s/day}) \quad (2.4)
\]

\[
\beta = 1/2 \cdot (1/\gamma - 1) + \omega \cdot (\text{timezone} - a \cdot \lambda) \cdot 3600 \text{s/hr} \quad (2.5)
\]

\[
\gamma = \arccos[-\tan(\pi \phi/180) \cdot \tan \delta] \quad (2.6)
\]

\[
\delta = 23.45\pi/180 \cdot \cos[2\pi/365 \cdot (172 - JD)] \quad (2.7)
\]

where \( q_{sw} \) is the mean short wave radiation flux [W/m\(^2\)] between time \( t_1 \) and \( t_2 \) in seconds, where \( t_{sunrise} < t_1, t_2 < t_{sunset} \) to avoid “negative” radiation, \( E_{sw} \) is the daily radiation energy [J/m\(^2\)], \( \omega \) is the frequency [s], \( \beta \) is a phase shift [rad] to adjust to the the local time, \( \gamma \) is the dimensionless photofraction, i.e. ratio between the duration the sun is above the horizon and the duration of a full day, \( \text{timezone} \) is the time difference [hr] between the local time imposed by law and UTC, \( a = 1/15 \) hr/deg, \( \lambda \) is the longitude [deg, positive to the east], \( \phi \) is the latitude [deg, pos. to the north], \( \delta \) is the declination [rad] of the sun at the equator and \( JD \) is the day of the year.

Equation 2.5 differs from the one of in the Science Manual Antenucci & Imerito (2001) in two respects: the second term was introduced to account for the 23 min difference between CET and mean solar time at Konstanz and in eq. 2.6 a minus sign had to be inserted, because this is incorrect in the manual.

Figure 2.6 shows, as an example, the radiation flux for three days without clouds for a time step of one hour.

**Relation between cloud cover fraction and radiation ratio** The last step in the estimation of SWR from CCF was to find, by means of a regression analysis, a relation between the radiation ratio and the CCF.

This was done in two ways. One method was to relate a CCF observation, which is an instantaneous value, with the measurement of SWR at the nearest hour, which is an hourly integral value. An alternative method was to take an average of all CCF values during daylight hours and the sum of all SWR measurements of that day.

With the first method, CCF observations before 9am or after 4pm were not considered, because corresponding calculated cloud-free SWR at dawn or dusk can be close to or equal to zero. This leads to extremely large or infinite values of the SWR ratio (i.e. measured SWR divided by calculated cloud-free SWR). With the second method, only CCF observations between 6am and 9pm were used for averaging, since night observations are of no interest.

Figure 2.7 shows results from the two approaches. The top and middle panels show the data pairs of SWR and CCF as measured at each weather station together with means and standard deviations. The means of SWR ratio in the right panels were calculated for 5% intervals of CCF at Güttingen and 2% intervals at Konstanz. The bottom panels show a
CHAPTER 2. METHODS

Figure 2.6: Mean hourly short wave radiation flux under cloud-free conditions calculated with eqs. 2.2–2.7 for Lake Constance

<table>
<thead>
<tr>
<th>location</th>
<th>instit.</th>
<th>from to</th>
<th>resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>Konstanz</td>
<td>DWD</td>
<td>01.01.60</td>
<td>31.12.00</td>
</tr>
<tr>
<td>Friedrichshafen</td>
<td>DWD</td>
<td>01.01.60</td>
<td>31.07.77</td>
</tr>
<tr>
<td>Eriskirch-Mariabrunn</td>
<td>DWD</td>
<td>01.08.80</td>
<td>31.05.86</td>
</tr>
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<td>Eriskirch-Mariabrunn</td>
<td>DWD</td>
<td>01.10.86</td>
<td>31.08.87</td>
</tr>
<tr>
<td>Eriskirch-Mariabrunn</td>
<td>DWD</td>
<td>01.12.87</td>
<td>31.12.00</td>
</tr>
<tr>
<td>Güttingen</td>
<td>MCH</td>
<td>01.01.81</td>
<td>31.12.96</td>
</tr>
</tbody>
</table>

Table 2.2: Overview of received cloud cover fraction data.

third order polynomial fit through the means of the combined data of both stations. The means are computed for 2 % intervals of CCF.

The polynomial fit with daily data appeared to be the better choice for the estimation of the missing SWR data as the means decrease in more monotonically with increasing CCF than with hourly values.

2.2.2 Cloud Cover Fraction

Neither one of the federal weather services measures long wave radiation near Lake Constance, but all of them observe cloud cover fraction. Cloud cover fraction (CCF) is simply the fraction of the sky that is covered by clouds (0 ≤ CCF ≤ 1). DYRESM will then calculate the incoming long wave radiation from CCF with eq. 1.8 on p. 14.

<table>
<thead>
<tr>
<th>from to</th>
<th>observ./day</th>
<th>observ. hours</th>
</tr>
</thead>
<tbody>
<tr>
<td>01/1960 06/1969</td>
<td>5</td>
<td>6:00, 9:00, ..., 18:00</td>
</tr>
<tr>
<td>07/1969 11/1972</td>
<td>6</td>
<td>6:00, 9:00, ..., 21:00</td>
</tr>
<tr>
<td>12/1972 12/1980</td>
<td>8</td>
<td>0:00, 3:00, ..., 21:00</td>
</tr>
<tr>
<td>01/1981 12/2000</td>
<td>24</td>
<td>0:00, 1:00, ..., 23:00</td>
</tr>
</tbody>
</table>

Table 2.3: Observation hours of cloud cover fraction at station Konstanz
Figure 2.7: Short wave radiation ratio (SWR ratio) vs. cloud cover fraction (CCF). a–c shows hourly values of SWR, d–f shows daily means of SWR and CCF. The squares indicate the mean at CCF, the triangles the distance of one standard deviation from the mean. For details see text.
CHAPTER 2. METHODS

Figure 2.8: Mean daily cloud cover fraction for each month and for the whole year (filled bars). Errorbars indicate the interval in which 90% of all values fall. Data from Deutscher Wetterdienst (DWD) at stations Konstanz and Friedrichshafen for a time period of 41 years. Only days with 8 or more observation were included.

Figure 2.9: Cloud cover fraction at hours with rainfall. Data from station Gütingen.

Weather stations, time periods and resolutions of obtained CCF observations are listed in table 2.2. CCF is not a continuously recorded variable, but rather observed several times a day. More than 99% of MCH’s records are at either 6:40, 12:40, 18:40 CET or 7:40, 13:40, 19:40 CET. Observations in Friedrichshafen are every 3 hours starting at midnight. Konstanz increased its observation frequency 3 times during the 41 year period (see table 2.3). In fig. 2.8 mean daily CCF is plotted for each month and for the whole year.

In order to come up with the 41 years hourly time series, the data were handled as follows: At each station except Eriskirch-Mariabrunn, the CCF values were interpolated to form a time series at full hours, whereby the maximum gap size of interpolation was 14 hours in order to fill-in the commonly missing night values. Then for each hour an average between the stations was calculated. Since the resulting time series still had large data gaps, the daily values from station Eriskirch-Mariabrunn were included to reduce the number of gaps to 145 with 2723 hours missing. An analysis of MCH’s data revealed that on hours where rainfall occurs, the corresponding CCF is on average 0.95 (fig. 2.9). Even after applying this relation, there were still 61 gaps with gap sizes of up to 38 hours and 1 gap with a gap size of 77 hours in the time series, which had to be linearly interpolated.
2.2. METEOROLOGICAL DATA

Figure 2.10: Mean daily air temperature for each month and the whole year (filled bars). Errorbars indicate the interval in which 90% of all values fall. Data from Deutscher Wetterdienst (DWD) at stations Konstanz (TU) and Friedrichshafen (TU) and from MeteoSwiss (MCH) at station Güttlingen.

Table 2.4: Overview of received air temperature data.

<table>
<thead>
<tr>
<th>location</th>
<th>instit.</th>
<th>from</th>
<th>to</th>
<th>resolution</th>
<th>average</th>
</tr>
</thead>
<tbody>
<tr>
<td>Konstanz (SY)</td>
<td>DWD</td>
<td>01.01.60</td>
<td>31.12.00</td>
<td>5-24 obs./day</td>
<td>9.6 °C</td>
</tr>
<tr>
<td>Konstanz (TU)</td>
<td>DWD</td>
<td>01.01.71</td>
<td>31.12.00</td>
<td>hourly</td>
<td>9.4 °C</td>
</tr>
<tr>
<td>Friedrichshafen (SY)</td>
<td>DWD</td>
<td>01.01.60</td>
<td>31.07.77</td>
<td>8 obs./day</td>
<td>9.1 °C</td>
</tr>
<tr>
<td>Friedrichshafen (TU)</td>
<td>DWD</td>
<td>01.01.65</td>
<td>31.07.77</td>
<td>hourly</td>
<td>8.9 °C</td>
</tr>
<tr>
<td>Eriskirch-Mariabrunn</td>
<td>DWD</td>
<td>01.08.80</td>
<td>31.05.86</td>
<td>daily</td>
<td></td>
</tr>
<tr>
<td>Eriskirch-Mariabrunn</td>
<td>DWD</td>
<td>01.10.86</td>
<td>31.08.87</td>
<td>daily</td>
<td></td>
</tr>
<tr>
<td>Eriskirch-Mariabrunn</td>
<td>DWD</td>
<td>01.12.87</td>
<td>31.12.00</td>
<td>daily</td>
<td></td>
</tr>
<tr>
<td>Güttlingen</td>
<td>MCH</td>
<td>01.01.81</td>
<td>31.12.00</td>
<td>hourly</td>
<td>9.1 °C</td>
</tr>
</tbody>
</table>

2.2.3 Air Temperature

Air temperature data were provided by DWD and MCH. Table 2.4 shows the details of the data. Konstanz and Friedrichshafen each have two partly overlapping but slightly different data sets (TU and SY). A systematic deviation was shown by a t-test which estimated for the overlapping time periods the 95% confidence intervals for the differences in the means as $-0.21 \pm 0.05$ °C for Konstanz and $-0.24 \pm 0.10$ °C for Friedrichshafen. The difference between the TU and the SY data sets may result from different measuring devices and from the different observation frequency. The observation hours of the SY data set are exactly the same as described above in the section on cloud cover. Figure 2.10 shows the daily means of air temperature for each month.

Measurements of the TU data were reported always 30 min after any full hour, measurements in Güttlingen 40 min after any full hour. In both cases the data were shifted forward by 30 min and 20 min respectively and not interpolated. Both, time shift and interpolation do conserve the mean, but interpolation reduces peak values. The full variability of the three parameters air temperature ($\theta_a$), vapour pressure ($e_a$) and wind speed ($U$), must be maintained. The reduction of variability that accompanies interpolation becomes important when two variables are multiplied as they are in the equations for
CHAPTER 2. METHODS

Table 2.5: Correction of wrong values in the SY data set of air temperature (°C) at Konstanz.

<table>
<thead>
<tr>
<th>time</th>
<th>original</th>
<th>corrected</th>
<th>comment</th>
</tr>
</thead>
<tbody>
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<td>08.07.69 18:00</td>
<td>-23</td>
<td>–</td>
<td>left empty</td>
</tr>
<tr>
<td>27.10.71 06:00</td>
<td>-17</td>
<td>7.5</td>
<td>from TU data set</td>
</tr>
<tr>
<td>17.05.86 18:00</td>
<td>-23.4</td>
<td>21.4</td>
<td>from TU data set</td>
</tr>
<tr>
<td>19.05.93 15:00</td>
<td>-20.3</td>
<td>–</td>
<td>left empty</td>
</tr>
</tbody>
</table>

Figure 2.11: Mean daily relative humidity for each month and the whole year (filled bars). Errorbars indicate the interval in which 90% of all values fall. Data from Deutscher Wetterdienst (DWD) at stations Konstanz (TU) and Friedrichshafen (TU) and from MeteoSwiss (MCH) at station Güttingen.

sensible \((U \cdot \theta_a)\) and latent \((U \cdot e_a)\) heat fluxes. This is because the mean of a product of two variables is not the same if the standard deviations of the variables are different even though their means are equal.

The number of data gaps in each data set was reduced by linear interpolation. The maximum gap size that was allowed to be interpolated was 1 hr in the TU data sets and Güttingen and 7 hr for the SY data sets. Temperature at station Eriskirch-Mariabrunn was not used at all, because the time series after interpolation between the other stations did not have data gaps larger than 8 hr.

The next step was to calculate a mean of all stations for each hour. It was assumed that the mean of all 5 sets (9.3 °C) is representative for the whole lake. Therefore each data set had to be corrected by a factor which would alter its mean to that of the whole lake to ensure that each data set would yield the representative mean if used alone.

The resulting temperature time series had 80 data gaps with a total of 526 missing hours, where the maximum gap size was 8 hr. A final linear interpolation was performed to fill-in these gaps.

2.2.4 Vapour Pressure

MeteoSwiss provided vapour pressure, while Deutscher Wetterdienst contributed relative humidity for all of its stations. Details are shown in table 2.6 and daily means of each month are shown in fig. 2.11.

Vapour pressure \(e_a\) (in mbar) and relative humidity \(h_r\) (in %) can easily be converted
2.2. METEOROLOGICAL DATA

<table>
<thead>
<tr>
<th>location</th>
<th>instit.</th>
<th>from</th>
<th>to</th>
<th>resolution</th>
<th>average</th>
</tr>
</thead>
<tbody>
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<td>Konstanz (TU)</td>
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<td>31.12.00</td>
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<td>78.8 %</td>
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<td>Friedrichshafen (TU)</td>
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<td>78.6 %</td>
</tr>
<tr>
<td>Eriskirch-Mariabrunn</td>
<td>DWD</td>
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<td>Eriskirch-Mariabrunn</td>
<td>DWD</td>
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<td>Eriskirch-Mariabrunn</td>
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<tr>
<td>Güttingen</td>
<td>MCH</td>
<td>01.01.81</td>
<td>31.12.00</td>
<td>hourly</td>
<td>78.9 %</td>
</tr>
</tbody>
</table>

Table 2.6: Overview of received vapour pressure or relative humidity data.

\[
h_r(JD) = A \cdot \sin[2\pi/T \cdot (JD - t_{sh})] + \text{mean}(h_r) \]

with \( A = 7.87 \% \), \( t_{sh} = 244.96 \text{ days} \), \( \text{mean}(h_r) = 78.86 \% \)

JD is the Julian Day of the year

\[ h_r(JD) = A \cdot \sin[2\pi/T \cdot (JD - t_{sh})] + \text{mean}(h_r) \] (2.9)

DWD’s records were measured 30 min, MCH’s records 40 min after any full hour. As with the air temperature, data were shifted forward in time for the same reasons. The number of data gaps of one missing value were reduced by linear interpolation. Records in Güttingen were converted into relative humidity and then an average of all stations was calculated. Since the means of relative humidity were within 0.3 % difference (see table 2.6), no correction towards an overall mean was done.

The resulting time series of relative humidity was complete after 01.01.1965, but the time period before was entirely missing. Since it is rather impossible to estimate humidity from other meteorological parameter, i.e. humidity is an independent variable, the slight periodic course of the monthly means was the justification to fit a sinus curve through the hourly data from 1965–2000:

\[
h_r(JD) = A \cdot \sin[2\pi/T \cdot (JD - t_{sh})] + \text{mean}(h_r) \] (2.9)

The least square fit resulted in an amplitude \( A \) of 7.87 %, a time shift \( t_{sh} \) of 244.96 days and a mean humidity \( \text{mean}(h_r) = 78.86 \% \), while the period was given as \( T = 365.25 \text{ days} \) and JD is the Julian Day. The root mean squared value was 14.84 %. As an example, one year of hourly data and the sinus curve of eq. 2.9 are plotted in fig. 2.12 for 1965.

Figure 2.12: Hourly data of relative humidity (crosses) and sinus curve fit (line) for year 1965. Data from Deutscher Wetterdienst (DWD) and MeteoSwiss (MCH).

\[
e_a(\theta_a) = \frac{h_r}{100} \cdot \exp[2.3026 \cdot \left(\frac{7.5 \cdot \theta_a}{\theta_a + 237.3} + 0.7858\right)]
\] (2.8)

Master’s Thesis Ralf Hornung
The University of Western Australia, Centre for Water Research
CHAPTER 2. METHODS

2.2.5 Wind Speed

Wind speed is recorded at Güttingen, Konstanz, Friedrichshafen and Eriskirch-Mariabrunn. Figure 2.13 shows mean daily wind speeds for each month. As with air temperature, two overlapping data sets (FF and SY) were available for Friedrichshafen and Konstanz (see table 2.7). FF data were reported at any full hour, while observation hours of the SY data set have already been described in sec. 2.2.2. The differences in the means between the FF and SY data for the overlapping periods are less than 0.03 m/s, i.e. there is no systematic difference.

The first step was to correct erroneous wind speed measurements (table 2.8). Due to their hourly resolution, the FF data sets were preferred and the SY data sets were only used to fill gaps in the FF data. Data from Eriskirch-Mariabrunn were not used at all

---

Table 2.7: Overview of received wind speed data.

<table>
<thead>
<tr>
<th>location</th>
<th>instit.</th>
<th>from</th>
<th>to</th>
<th>resolution</th>
<th>average m/s</th>
</tr>
</thead>
<tbody>
<tr>
<td>Konstanz (FF)</td>
<td>DWD</td>
<td>01.01.60</td>
<td>31.12.00</td>
<td>hourly</td>
<td>1.94</td>
</tr>
<tr>
<td>Konstanz (SY)</td>
<td>DWD</td>
<td>01.01.60</td>
<td>31.12.00</td>
<td>5-24 obs./day</td>
<td>1.93</td>
</tr>
<tr>
<td>Friedrichshafen (FF)</td>
<td>DWD</td>
<td>01.01.65</td>
<td>31.07.77</td>
<td>hourly</td>
<td></td>
</tr>
<tr>
<td>Friedrichshafen (FF)</td>
<td>DWD</td>
<td>01.09.88</td>
<td>28.02.89</td>
<td>hourly</td>
<td></td>
</tr>
<tr>
<td>Friedrichshafen (FF)</td>
<td>DWD</td>
<td>01.05.89</td>
<td>28.02.99</td>
<td>hourly</td>
<td></td>
</tr>
<tr>
<td>Friedrichshafen (FF)</td>
<td>DWD</td>
<td>07.11.00</td>
<td>31.12.00</td>
<td>hourly</td>
<td>2.98 (1965-2000)</td>
</tr>
<tr>
<td>Friedrichshafen (SY)</td>
<td>DWD</td>
<td>01.01.65</td>
<td>31.07.77</td>
<td>8 obs./day</td>
<td>2.63</td>
</tr>
<tr>
<td>Eriskirch-Mariabrunn</td>
<td>DWD</td>
<td>01.08.80</td>
<td>31.05.86</td>
<td>daily</td>
<td></td>
</tr>
<tr>
<td>Eriskirch-Mariabrunn</td>
<td>DWD</td>
<td>01.10.86</td>
<td>31.08.87</td>
<td>daily</td>
<td></td>
</tr>
<tr>
<td>Eriskirch-Mariabrunn</td>
<td>DWD</td>
<td>01.12.87</td>
<td>31.12.00</td>
<td>daily</td>
<td></td>
</tr>
<tr>
<td>Güttingen</td>
<td>MCH</td>
<td>01.01.81</td>
<td>31.12.00</td>
<td>hourly</td>
<td>2.19</td>
</tr>
</tbody>
</table>
2.2. METEOROLOGICAL DATA

<table>
<thead>
<tr>
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<th>time</th>
<th>original</th>
<th>corr.</th>
<th>comment</th>
</tr>
</thead>
<tbody>
<tr>
<td>FN (SY)</td>
<td>03.12.65 03:00</td>
<td>49.9</td>
<td>2.5</td>
<td>from FF data set</td>
</tr>
<tr>
<td>FN (FF)</td>
<td>26.08.96 12:00, 13:00</td>
<td>18.9, 30.8</td>
<td></td>
<td>left empty</td>
</tr>
<tr>
<td>KN (FF)</td>
<td>02.02 07:00-03.02.94 11:00</td>
<td>up to 19</td>
<td></td>
<td>other stations hardly any wind</td>
</tr>
</tbody>
</table>

Table 2.8: Correction of wrong values of wind speed (m/s) at Konstanz (KN) and Friedrichshafen (FN).

<table>
<thead>
<tr>
<th>location</th>
<th>instit.</th>
<th>from</th>
<th>to</th>
<th>res.</th>
<th>average mm/yr</th>
</tr>
</thead>
<tbody>
<tr>
<td>Konstanz</td>
<td>DWD</td>
<td>01.01.60</td>
<td>31.12.00</td>
<td>daily</td>
<td>849</td>
</tr>
<tr>
<td>Friedrichshafen</td>
<td>DWD</td>
<td>01.01.60</td>
<td>30.11.88</td>
<td>daily</td>
<td>997</td>
</tr>
<tr>
<td>Eriskirch-Mariabrunn</td>
<td>DWD</td>
<td>01.08.80</td>
<td>31.05.86</td>
<td>daily</td>
<td></td>
</tr>
<tr>
<td>Eriskirch-Mariabrunn</td>
<td>DWD</td>
<td>01.10.86</td>
<td>31.08.87</td>
<td>daily</td>
<td></td>
</tr>
<tr>
<td>Güttingen</td>
<td>MCH</td>
<td>01.01.81</td>
<td>31.12.00</td>
<td>hourly</td>
<td>953</td>
</tr>
</tbody>
</table>

Table 2.9: Overview of received precipitation data.

because of the daily resolution.

The wind stress is proportional to the wind speed squared, while the equations for sensible and latent heat are proportional to the wind speed. Prior to averaging of the stations a representative mean wind speed of the whole lake had to be computed. It was considered more important to find a wind speed which represents the average wind stress than the average heat transfers. Hence, for each of the three time series at Konstanz, Friedrichshafen and Güttingen, a mean of wind speed squared was calculated:

Konstanz 6.4 (m/s)^2
Güttingen 7.4 (m/s)^2
Friedrichshafen 13.6 (m/s)^2

It was then assumed that a mean squared value of 10 (m/s)^2 might represent the spatial average over the whole lake. The underlying idea was that the mean between Güttingen and Konstanz of 6.9 (m/s)^2 may represent the wind stress in the southwest corner of the lake and the value for Friedrichshafen may represent the northeast corner.

Each time series was then corrected by a multiplication factor so that all of them had a new average of 10 (m/s)^2. The average of all stations formed the final wind speed time series.

It must be noted that this wind speed time series is different from one which would have been calculated with non-squared values.

### 2.2.6 Precipitation

Precipitation rate, which is height of precipitation per unit time, was only requested in daily resolution. The obtained data are listed in table 2.9.

Measurements at stations of DWD were always recorded at 7.30am and referred to the precipitation height of the previous 24 hours. DWD’s values were therefore partitioned such that 7.5/24 were assigned to the day where this value was reported and the remaining 16.5/24 to the day before. Then the daily precipitation rate was divided by 24 to create an hourly time series.
CHAPTER 2. METHODS

The mean precipitation increases along the longitudinal axes of the lake from north west to south east considerably (Bauerle et al., 1998). Wagner et al. (1994) assume a mean precipitation rate of 0.449 km$^3$/yr, which equals 1002 mm/yr. All time series were therefore multiplied by a factor so that each had a new mean of 1002 mm/yr and then an average of each hour was taken. In the resulting time series one gap of 24 hr on 31.12.70 was linearly interpolated. Daily means of precipitation are plotted in fig. 2.14.

2.3 Inflow Data

For each inflow, river or groundwater, DYRESM requires a time series in daily resolution of discharge volume [m$^3$], mean water temperature [°C] and mean salinity [pss].

The following institutions are responsible for flow gauging and sampling in the states of Germany, Switzerland and Austria respectively:

- The State Institute for Environmental Protection Baden-Württemberg (Landesanstalt für Umweltschutz LiU) with the departments Hochwasser-Vorhersage-Zentrale (HVZ), which measures flow, and the Institut für Seenforschung (ISF), which stores data of measuring campaigns of the Internationale Gewässerschutzkommission für den Bodensee (IGKB).
- The Federal Office for Water and Geology FOWG (Bundesamt für Wasser und Geologie BWG)
- The Hydrographischer Dienst Vorarlberg at the Landeswasserbauamt Bregenz (LWBA) for discharge and water temperature measurements and the Umweltinstitut des Landes Vorarlberg (here abbreviated as UILV) for sampling of chemical parameters.

These institutions were asked to provide discharge $Q$, temperature $\theta$ and salinity or conductivity $\kappa$ for all gauged inflowing rivers in daily resolution for the time period 1960–2000. As a result data for 13 rivers were supplied, which will be described in the order of their mean discharge in the following sectionsec. 2.3.2–2.3.12. Finally, a 14th river “Residual Flow” needed to be introduced to represent all ungauged rivers (sec. 2.3.13). The prepared

Since all institutions measure electric(al) conductivity (e. g. in units of $\mu S/cm$), sec. 2.3.1 explains how conductivity was converted to salinity.

### 2.3.1 Conversion of Electrical Conductivity into Salinity

DYRESM calculates density from temperature, salinity and pressure with the UNESCO International Equation of State (IES) for Seawater (Antenucci & Imerito, 2001).

In the definition of the Practical Salinity Scale PSS78, salinity $S$ of a sample of sea water is defined as some function of the ratio $X$ of the electrical conductivity at 15 °C and one standard atmosphere, to that of a potassium chloride (KCl) standard solution, in which the mass fraction of KCl is 0.0324356 at the same temperature and pressure. A conductivity value of $\kappa_0 = 4.29142 \, S/m$ can be used for the KCl standard. When the ratio $X$ equals unity, practical salinity equals by definition 35 (Fofono, 1985). Salinity has no units, but is often given the “units” PSS78 in reference to the above definition.

Conductivity at a given temperature depends on the concentration of ions and the ion species distribution, which is different in seawater and freshwater. Therefore, the above definition of salinity cannot be applied to calculate salinity from conductivity measurements in freshwater. Bauerle et al. (1998) suggest to calculate the density $\rho_{LC}$ of freshwater in Lake Constance at ambient pressure as

$$
\rho_{LC}(\theta_w, \kappa_{20}) = \rho_{LC}(\theta_w) + C_\kappa \cdot \kappa_{20} \quad (2.10)
$$

$$
\rho_{LC}(\theta_w) = a_0 + a_1 \cdot \theta_w + a_2 \cdot \theta_w^2 + a_3 \cdot \theta_w^3 \quad (2.11)
$$

where $\theta_w$ is the water temperature, $\kappa_{20}$ is the conductivity normalised to 20 °C in $\mu S/cm$ and Heinz (1990) reports for $C_\kappa$ a value of $0.67 \cdot 10^{-3} \, \text{kg m}^{-3}/(\mu \text{S cm}^{-1})$, which is based on the assumption that the three major salts Ca(HCO$_3$)$_2$, MgSO$_4$ and NaCl are present in a ratio of 7:2:1. The coefficients $a_i$ are: $a_0 = 999.8429 \, \text{kg/m}^3$, $a_1 = 6.54891 \cdot 10^{-2} \, \text{kg/(m}^3 \text{ °C})$, $a_2 = 8.56272 \cdot 10^{-3} \, \text{kg/(m}^3 \text{ °C}^2)$, $a_3 = 5.9385 \cdot 10^{-5} \, \text{kg/(m}^3 \text{ °C}^3)$.

Any conversion must assure that a density $\rho_{LC}$ calculated with 2.10 is equal to a density $\rho_{IES}$ calculated with the UNESCO equation and the converted “density equivalent” salinity. To achieve this, the two density equations were equated and then rearranged to solve for conductivity

$$
\kappa_{20} = \left[ \rho_{IES}(\theta_w, S) - \rho_{LC}(\theta_w) \right]/C_\kappa \quad (2.12)
$$

This equation was evaluated for a range of expected temperature and salinity values ($\theta = 0\ldots25 \, ^\circ \text{C}$, $S = 0.1\ldots0.5$) and the resulting pairs of conductivity and equivalent salinity are plotted in fig. 2.15. The conversion equation was found by linear regression

$$
S = 8.50 \cdot 10^{-4} \, \text{cm} \, \mu \text{S}^{-1} \cdot \kappa_{20} - 4.72 \cdot 10^{-4} \quad (2.13)
$$

### 2.3.2 Alpine Rhine

Data for the Alpine Rhine were supplied by FOWG, LWBA and UILV. The locations, time periods and resolutions of discharge, temperature and conductivity are listed in table 2.10. In fig. 2.16, means and the intervals into which 90 % of all values fall are plotted for each month.
CHAPTER 2. METHODS

Figure 2.15: “Density equivalent” seawater salinity versus electrical conductivity at Lake Constance. See text for details.

<table>
<thead>
<tr>
<th>type</th>
<th>location</th>
<th>instit.</th>
<th>from</th>
<th>to</th>
<th>resolution</th>
<th>average</th>
</tr>
</thead>
<tbody>
<tr>
<td>Q</td>
<td>Schmitter</td>
<td>FOWG</td>
<td>01.01.60</td>
<td>31.12.83</td>
<td>daily</td>
<td>234.9 m³/s</td>
</tr>
<tr>
<td>Q</td>
<td>Diepoldsau</td>
<td>FOWG</td>
<td>01.01.84</td>
<td>31.12.00</td>
<td>daily</td>
<td>237.4 m³/s</td>
</tr>
<tr>
<td>Q</td>
<td>Lustenau</td>
<td>LWBA</td>
<td>02.01.76</td>
<td>31.12.00</td>
<td>daily</td>
<td>236.0 m³/s</td>
</tr>
<tr>
<td>θ</td>
<td>Schmitter</td>
<td>FOWG</td>
<td>19.06.62</td>
<td>30.06.66</td>
<td>daily</td>
<td>7.7 °C</td>
</tr>
<tr>
<td>θ</td>
<td>Diepoldsau</td>
<td>FOWG</td>
<td>01.03.69</td>
<td>31.12.00</td>
<td>daily</td>
<td>7.8 °C</td>
</tr>
<tr>
<td>θ</td>
<td>Lustenau</td>
<td>LWBA</td>
<td>01.01.90</td>
<td>31.12.00</td>
<td>daily</td>
<td>7.7 °C</td>
</tr>
<tr>
<td>κ</td>
<td>Schmitter</td>
<td>FOWG</td>
<td>01.01.76</td>
<td>31.12.83</td>
<td>daily</td>
<td>255 μS/cm</td>
</tr>
<tr>
<td>κ</td>
<td>Diepoldsau</td>
<td>FOWG</td>
<td>01.01.84</td>
<td>31.12.00</td>
<td>daily</td>
<td>260 μS/cm</td>
</tr>
<tr>
<td>κ</td>
<td>Fußach</td>
<td>UIVL</td>
<td>01.01.89</td>
<td>31.12.00</td>
<td>1 meas./3 weeks</td>
<td>305 μS/cm</td>
</tr>
</tbody>
</table>

Table 2.10: Overview of received inflow data for Alpine Rhine.

Figure 2.16: Mean daily discharge, temperature and electrical conductivity of Alpine Rhine for each month (crosses with connecting lines). The bars span the interval in which 90 % of all values fall into. “n” is the number of values for each statistic.
2.3. INFLOW DATA

The discharge time series was created by averaging the discharge measurements of FOWG and LWBA from 02.01.1976 on and taking the FOWG values before. The resulting time series was free of data gaps.

On days where two institutions had a temperature measurement, the average was taken. For the data gap from 30.06.1966 to 01.03.1969 and for the time before 19.06.1962 the river temperature was estimated by linear regression with air temperature. The best regression result was with a 3-day average of the air temperature including the current day (see fig. 2.17). Prediction of water temperatures below 0 °C were set to 0 °C. Otherwise there were no data gaps.

As with temperature, an average salinity was calculated when two institutions provided a value. Linear interpolation was done on 21 data gaps that were smaller than 7 days. For larger data gaps and for the time before 1976 monthly means were assumed. Salinity could not be correlated to discharge even if each month was treated separately. Conductivity was converted to salinity according to eq. 2.13.

2.3.3 Bregenzer Ach

Data for this river came from LWBA and UILV. The details are listed in table 2.11.

The discharge time series had three gaps of one day which were linearly interpolated.

Six temperature data gaps of one day and one data gap of two days were linearly interpolated. The temperatures from 19.06.1962–31.12.1978 were estimated by linear regression with the Alpine Rhine temperature (from FOWG measurements). The correlation was excellent which can be explained by the similar environmental conditions in both catchments (see fig. 2.19). For the time period before 19.06.1962 regression with average air temperature over 5 days was employed to estimate river water temperature (see fig. 2.20).
CHAPTER 2. METHODS

Figure 2.18: Mean daily discharge, temperature and electrical conductivity of Bregenzer Ach for each month (crosses with connecting lines). The bars span the interval in which 90% of all values fall into. “n” is the number of values for each statistic.

<table>
<thead>
<tr>
<th>type</th>
<th>location</th>
<th>instit.</th>
<th>from</th>
<th>to</th>
<th>resolution</th>
<th>average</th>
</tr>
</thead>
<tbody>
<tr>
<td>Q</td>
<td>Kennelbach</td>
<td>LWBA</td>
<td>02.01.76</td>
<td>31.12.00</td>
<td>daily</td>
<td>46.5 m³/s</td>
</tr>
<tr>
<td>θ</td>
<td>Kennelbach</td>
<td>LWBA</td>
<td>01.01.79</td>
<td>31.12.00</td>
<td>daily</td>
<td>7.6 °C</td>
</tr>
<tr>
<td>κ</td>
<td>Bregenz</td>
<td>UILV</td>
<td>01.01.89</td>
<td>31.12.00</td>
<td>1 meas./3 weeks</td>
<td>300 µS/cm</td>
</tr>
</tbody>
</table>

Table 2.11: Overview of received inflow data for Bregenzer Ach.

Figure 2.19: Linear regression of daily water temperature of Bregenzer Ach and Alpine Rhein.
2.3. INFLOW DATA

Figure 2.20: Linear regression of water temperature of Bregenzer Ach with the average air temperature of n-days (including the current day).

<table>
<thead>
<tr>
<th>type</th>
<th>location</th>
<th>instit.</th>
<th>time period</th>
<th>resolution</th>
<th>average</th>
</tr>
</thead>
<tbody>
<tr>
<td>Q</td>
<td>Gießen</td>
<td>LfU (HVZ)</td>
<td>01.01.60–31.12.00</td>
<td>daily</td>
<td>19.7 m³/s</td>
</tr>
<tr>
<td>θ</td>
<td>unknown</td>
<td>LfU (ISF)</td>
<td>1978/79</td>
<td>34 samples</td>
<td>7.5 °C</td>
</tr>
<tr>
<td>κ</td>
<td>unknown</td>
<td>LfU (ISF)</td>
<td>1978/79, 85/86, 95–97</td>
<td>268 samples</td>
<td>401 μS/cm</td>
</tr>
</tbody>
</table>

Table 2.12: Overview of received inflow data for Argen.

In cases where the regression predicted temperatures below 0 °C, the value was set to 0 °C.

All data gaps in the salinity time series were filled with monthly means. Interpolation was not considered, because measurements were collected, on average, every 3 weeks and any extreme values would then have an influence for many days.

2.3.4 Argen

The details of the data for river Argen are listed in table 2.12.

The discharge data were complete.

All missing temperature values were estimated by linear regression between water and average air temperature over three days (see fig. 2.21) and subsequent correction of values below 0 °C.

The salinity time series was created by using monthly means for all data gaps.
Figure 2.21: Linear regression of water temperature of Argen with the average air temperature of n-days (including the current day).

<table>
<thead>
<tr>
<th>type</th>
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<th>instit.</th>
<th>time period</th>
<th>resolution</th>
<th>average</th>
</tr>
</thead>
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<tr>
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<td>Gebertshaus</td>
<td>LfU (HVZ)</td>
<td>01.01.60–31.12.00</td>
<td>daily</td>
<td>11.2 m³/s</td>
</tr>
<tr>
<td>θ</td>
<td>unknown</td>
<td>LfU (ISF)</td>
<td>1978/79</td>
<td>28 samples</td>
<td>8.2 °C</td>
</tr>
<tr>
<td>κ</td>
<td>unknown</td>
<td>LfU (ISF)</td>
<td>1978/79, 85/86, 95–97</td>
<td>307 samples</td>
<td>572 μS/cm</td>
</tr>
</tbody>
</table>

Table 2.13: Overview of received inflow data for Schussen.

2.3.5 Binnenkanal

Only discharge data were provided by FOWG for the Binnenkanal, or more precise the Rheintal-Binnenkanal, at St.Margarethen for the period 1960–2000 in daily resolution (mean discharge 12.5 m³/s ). The Binnenkanal is a channel situated to the left of Alpine Rhine and running parallel to Alpine Rhine for about 20-30 km. It seems to collect creeks that would otherwise flow into Alpine Rhine. Temperature and salinity were assumed to be the same as in Alpine Rhine.

2.3.6 Schussen

The details of the data for river Schussen are listed in table 2.13.

The discharge time series was complete.

The data gaps in the water temperature time series were estimated by linear regression with average air temperature. A 3 day average yielded the best fit (see fig. 2.22, panel A). Predictions below 0 °C were set to 0 °C.

The salinity time series on days without measurements consists of monthly means which were calculated from the 307 samples of conductivity.
Figure 2.22: Linear regression of river water temperature with the average air temperature of n-days (including the current day) for 6 different rivers. For each river n was varied between 2 and 5 days, but only the best of the 4 regressions is shown.
CHAPTER 2. METHODS

<table>
<thead>
<tr>
<th>type</th>
<th>location</th>
<th>instit.</th>
<th>from</th>
<th>to</th>
<th>resolution</th>
<th>average</th>
</tr>
</thead>
<tbody>
<tr>
<td>Q</td>
<td>Lauterach</td>
<td>LWBA</td>
<td>01.01.84</td>
<td>31.12.00</td>
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<td>6.9 m³/s</td>
</tr>
<tr>
<td>θ</td>
<td>Lauterach</td>
<td>LWBA</td>
<td>19.01.89</td>
<td>31.12.00</td>
<td>ca. biweekly</td>
<td>9.4 °C</td>
</tr>
<tr>
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<td>Lauterach</td>
<td>UILV</td>
<td>19.01.89</td>
<td>31.12.00</td>
<td>ca. biweekly</td>
<td>573 µS/cm</td>
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Table 2.14: Overview of received inflow data for Dornbirner Ach.

Figure 2.23: Linear regressions of discharge (left panel) and water temperature (right panel) of Dornbirner Ach and Bregenzer Ach.

### 2.3.7 Dornbirner Ach

Flow measurements for this river are carried out by LWBA while sampling is done by UILV (see table 2.14).

In the discharge data four data gaps smaller than 2 days had to be linearly interpolated. Discharge before 1984 was computed by linear regression with Bregenzer Ach, which is the neighbouring catchment and also located in an alpine environment (see fig. 2.23, left panel).

In the salinity data sixteen data gaps smaller than 3 days were linearly interpolated. For the time period in which temperature measurements at Bregenzer Ach were recorded, i.e. after 1979, water temperatures at Dornbirner Ach were calculated with a linear regression from Bregenzer Ach (see fig. 2.23, right panel). Before 1979, and for all remaining data gaps, the time series was completed by means of linear regression with average air temperature over 4 days (see fig. 2.22, panel B).

Monthly means of salinity were used where data were missing.

### 2.3.8 Leiblach

As with Dornbirner Ach, LWBA and UILV are responsible for gauging and sampling (see table 2.15).

The discharge time series had no data gaps in the provided period. River Argen drains the area northwest of the catchment of Leiblach. Therefore discharge of Leiblach before...
2.3. INFLOW DATA

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<th>to</th>
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<td>LWBA</td>
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<td>UILV</td>
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<td>UILV</td>
<td>23.08.89</td>
<td>04.12.00</td>
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<td>440 $\mu$S/cm</td>
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</table>

Table 2.15: Overview of received inflow data for Leiblach.

![Graph showing discharge relations](image)

Figure 2.24: Linear regression of discharge of Leiblach and Argen (left panel) and discharge of Seefelder Aach and Rotach (right panel).

1977 was estimated by linear regression with discharge of Argen (see fig. 2.24, left panel).

The temperature data were linearly interpolated for data gaps equal to or smaller than 2 days. The remaining missing values were estimated by linear regression with average air temperature over 4 days (see fig. 2.22, panel C).

The salinity time series was created by filling data gaps with monthly means.

2.3.9 Seefelder Aach

Details of data for this river are listed in table 2.16.

The number of data gaps of up to 5 days in the discharge data was reduced from 138 to 69 by linear interpolation. The remaining data gaps were filled by linear regression with river Rotach (see fig. 2.24, right panel).

The 24 samples of water temperature were used in a linear regression with average air

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Table 2.16: Overview of received inflow data for Seefelder Aach.
CHAPTER 2. METHODS

<table>
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Table 2.17: Overview of received inflow data for Rotach.

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<td>daily</td>
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<td>1978/79, 85/86, 95–97</td>
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Table 2.18: Overview of received inflow data for Stockacher Aach.

temperature over 2 days. This regression was then applied to create the 41 years time series of water temperature (see fig. 2.22, panel D).

Monthly means of salinity were used for all missing values.

2.3.10 Rotach

The HVZ and the ISF supplied data for this river, where details are given in table 2.17.

The discharge records were without any data gaps.

The data gaps in the temperature records were filled using a linear regression with average air temperature over 3 days (see fig. 2.22, panel E).

The salinity time series consists of monthly means where data were missing.

2.3.11 Stockacher Aach

Details of the received data for Stockacher Aach can be found in table 2.18.

The flow gauging takes place at a diverted canal and at the river itself. For days on which no records were available, the discharge was calculated by linear regression from Seefelder Aach. Seefelder Aach is the closest river of similar size and has the same kind of catchment. For reasons of simplicity, this relationship was also used for the few days where only either a river discharge or a canal discharge was reported, even though a canal-river regression could have been employed instead. Due to a data gap in the Seefelder Aach data, a 59-day period could not be calculated. A linear regression with the river Rotach was then used to compute values for these 59 days (see fig. 2.25).

All water temperatures except the 24 days in 1978/79 were estimated by means of a linear regression with average air temperature over 2 days (see fig. 2.22, panel F).

Monthly means of salinity were used for all days except the ones on which samples were taken.
2.3. INFLOW DATA

2.3.12 Goldach, Steinach and Aach

The FOWG provided discharge records for these three minor rivers entering the lake on the south shore (see table 2.19). No temperature or salinity data were available.

There were no data gaps in the time periods listed in table 2.19. However, since no neighbouring rivers were available for a regression analysis, monthly means of discharge had to be used for the time period before 1969. The discharge of Goldach between 1969 and 1974 was computed by linear regression from Steinach, which has the neighbouring catchment to Goldach.

River Leiblach was assumed to be the most similar river to Goldach, Steinach and Aach in terms of the heat budget as these catchments reach elevations up to around 800–1000 m a.s.l. and the mean discharges are of comparable order. Furthermore Leiblach has the largest number of temperature measurements of the smaller rivers. Therefore the water to air temperature relation of Leiblach was used for these three rivers as well (see fig. 2.22, panel C).

For the same reasons as with temperature, the monthly means of salinity of Leiblach were applied to the three Swiss rivers.

2.3.13 Inflow Correction

There are many creeks flowing into the lake for which no data exist. Furthermore the nature of discharge measurements leads to values of rather high uncertainty.
CHAPTER 2. METHODS

Figure 2.26: Comparison of simulated water level with field data. Inflow data includes an extra inflow according to eq. 2.14, but no correction according to eq. 2.16.

At first it was thought that the errors in the discharge measurements were random and sufficiently small, so that the total annual mean inflow $Q_{res} = 9.3 \text{ m}^3/\text{s}$ of the ungauged creeks can be approximated by a simple water balance of annual means:

$$Q_{res} = Q_{Rhine, out} + Q_{eva} + Q_{wdr} - Q_{precip} - Q_{inf}$$  \hspace{1cm} (2.14)

$$Q_{inf} = \sum_{i=1}^{13} Q_i$$  \hspace{1cm} (2.15)

with the discharge of the out-flowing Rhine $Q_{Rhine, out} = 354.8 \text{ m}^3/\text{s}$ (see sec. 2.4), the mean evaporation $Q_{eva} = 9.0 \text{ m}^3/\text{s}$ (Wagner et al., 1994), the total amount of withdrawals $Q_{wdr} = 5.55 \text{ m}^3/\text{s}$ (Stabel & Kleiner, 1995), the direct precipitation onto the lake $Q_{precip} = 14.2 \text{ m}^3/\text{s}$ (Wagner et al., 1994) and the sum of all 13 gauged rivers $Q_{inf}$ as 345.8 $\text{m}^3/\text{s}$ (see sec. 2.3.2–2.3.12 and mean of Alpine Rhine is 235.9 $\text{m}^3/\text{s}$ from FOWG data).

For any given day, the discharge of the ungauged creeks $Q_{res,daily}$ was computed relative to the discharge $Q_{big5,daily}$ of the 5 largest rivers (Alpine Rhine, Bregenzer Ach, Argen, Binnenkanal and Schussen): $Q_{res,daily} = Q_{res}/Q_{big5} \cdot Q_{big5,daily}$, where the fraction $Q_{res}/Q_{big5}$ is based on the annual means.

However, model runs over periods of 5–10 years showed water level deviations of several metres between field data and simulation. Figure 2.26 shows, as an example, a 9 year simulation. In general the simulated water levels tended to continuously increase for model runs later than mid the 70s, while the opposite happened before.

The yearly sum of evaporation, withdrawal and precipitation is approximately zero and yearly variations of these three variables can only cause this sum to have a value of a few cubic metres. For short periods, differences between simulation and model will also be due to inaccurate knowledge of the lake’s geometry. But for time periods longer than a full cycle of the lake water level, i.e. the water level is back at the initial height, the inaccuracy does not matter at all. This is because after reaching the initial height again, the change in lake volume is zero, which also must be the case for the model independent of the assumed geometry. Thus the deviation of nearly 3 m at the beginning of 1998,
2.3. INFLOW DATA

when the lake level is close to the one at the beginning of 1990, must be almost entirely attributed to wrong inflow and outflow data.

An underestimation of inflows for the period until the mid 70s, and an overestimation of inflows after the mid 70s must have occurred (or vice versa for the outflow). The magnitude of these errors was so large that the calculation of residual flow above became obsolete.

In order to be able to model a few decades of stratification without running into trouble because of extreme lows or highs in water levels, the inflow and outflow data needed to be corrected from field data of water levels. The required additional discharge $Q_{corr}$ is:

$$Q_{corr}(t_1 \leq t \leq t_2) = \frac{1}{t_2 - t_1} \cdot [V(t_2) - V(t_1) - \int_{t_1}^{t_2} (Q_{inf}(t) - Q_{Rhine;out}(t)) dt] \hspace{1cm} (2.16)$$

where $t_1$ and $t_2$ are always the first and last day of a month, respectively, $V$ is the lake volume calculated from the water level elevation on that day, $Q_{inf}$ is the sum of the discharge of all 13 inflowing rivers and $Q_{Rhine;out}$ is the discharge of the outflowing Seerhein.

To summarise the above correction procedure, the additional discharge $Q_{corr}$ can be interpreted as:

$$Q_{corr} = \epsilon_{inf} + \epsilon_{Rhine;out} + \zeta_{eva} + \zeta_{wdr} + \zeta_{precip} \hspace{1cm} (2.17)$$

where $\epsilon$ terms stand for the errors in the inflow and outflow measurements, $\zeta$ terms stand for the deviation of this variable from the long term average during the month for which the $Q_{corr}$ was calculated and indices are the same as before.

At times where $Q_{corr}$ was positive, it became an additional inflow. Otherwise it was included as an additional surface outflow. In fig. 2.27 $Q_{corr}$ is plotted versus time.

Temperature and salinity were assumed to be the same as river Leiblach, which might be representative for the numerous small creeks.
CHAPTER 2. METHODS

2.4 Outflow Data

DYRESM’s “Withdrawal File” contains discharge time series of all outflows and withdrawals in daily resolution. Lake Constance’s natural outflow is the Seerhein at Konstanz (sec. 2.4.1). Furthermore there are numerous drinking water withdrawals that were combined into one withdrawal (sec. 2.4.2). Additional to these a further outflow was required to correct the measured inflow and outflow data 2.4.3. The prepared withdrawals file for Lake Constance starts on 01.01.1960 and ends on 31.12.2000 (14976 points in time).

2.4.1 Seerhein

Almost 17 years (01.01.85–17.10.01) of daily discharge were provided by The State Institute for Environmental Protection Baden-Württemberg (LfU). The mean discharge for this period is 347.8 m$^3$/s. Because discharge can be expected to be closely related to water levels, the time period 1960–1984 was calculated by means of a least square fit.

Wagner et al. (1994) fitted a function of the form $Q = a_1 + a_2 H + a_3 H/(a_4 + \exp(-H/100))$ to 16 discharge measurements to allow calculation of discharge Q from water level readings H of level metre Konstanz, where $a_1 \ldots a_4$ are fitting parameters. The 16 data pairs (black squares) and the suggested fit (black dashed line) are plotted in fig. 2.28.

This was compared with a least square fit of the same functional form as above, but with the discharge and water level data provided by LfU, which formed 6134 data pairs. The data pairs are plotted in the same figure as grey crosses and the resulting fit as a dark grey line. Fitting coefficients and the RMS value can be found in the legend.

The water levels from table 2 in Wagner et al. (1994) coincide with the ones provided by LfU for this thesis (see also sec. 2.7), which means the same water level gauge is used. However, the corresponding discharges do not coincide, even though they also used data from LfU. It is common to calculate discharges from water levels with some function and to adjust this relation periodically with actual discharge measurements (propeller profiles,
It is presumed that Wagner et al. (1994) were given these actual discharge values, whereas the data for this thesis is based on the regularly updated discharge-water level function.

With this presumption, the 41 year outflow time series could then either be completely calculated from water levels with the fit given by Wagner et al. (1994) or alternatively the 17 years of discharge from LfU are used and the missing period before is calculated from the fit on that data. The latter was preferred.

2.4.2 Withdrawals

The sum of all withdrawals is 5.55 m³/s or 175 million m³/yr according to Stabel & Kleiner (1995) during the time of their publication. In this thesis it was not only assumed that this amount was constant during the 41 year period, but also that there is neither a seasonal nor a daily variation.

The long term constancy might be justified with the fact that water consumption per capita has dropped and might have compensated for the increase in population. This is at least confirmed for the period 1984–1996 where total withdrawal was between 165 and 185 million m³/yr (LfU, 1997, p. 136).

Relative variations within a year are of the order of +50% to −10% around the mean for water works of this size (Mutschmann & Stimmelmayr, 1995, p. 19). The maximum absolute deviation (+2.8 m³/s) from the mean is therefore negligible compared to magnitude of the errors involved in the stream gauging.

2.4.3 Outflow Correction

This discharge time series needed to be introduced to correct errors in the inflow and outflow data. See eq. 2.16 in sec. 2.3.13.

2.5 Parameters

This section is about the parameter settings which were used as default values. These parameters are set in two different files, the “Parameters File” and the “Configuration File”.

Some parameters can be considered as generic constants, i.e. they appear to be independent of any other environmental variable. All parameters that belong to that category are listed (among others) in the parameters-file and will be described in sec. 2.5.1. Most settings typically required for a numerical model (e.g. time step) can be found in the configuration-file. However, two more settings of that kind (output time and critical wind speed) are listed in the parameters file. They are briefly discussed in sec. 2.5.2. The only truly lake specific parameter is the annual mean, depth averaged light extinction coefficient (sec. 2.5.3).

2.5.1 Generic Constants

An overview of the generic constants used is provided in table 2.20.

2.5.1.1 Bulk Aerodynamic Transfer Coefficients $C_M$, $C_S$ and $C_L$

The coefficient $C_M$ is required to calculate the wind stress exerted on the water surface from air density and wind speed as given in eq. 1.11 on p. 15. $C_M$ is a weak function of wind speed, fetch length, wind duration, water depth (influences the surface waves) and a strong function of atmospheric stability above the water surface (Fischer et al., 1979). $C_M$
The value of $C_M$ is found in an iterative procedure as outlined in Imberger & Patterson (1990). In this thesis, however, neutral atmospheric stability was assumed and the value of $1.3 \cdot 10^{-3}$ as suggested by Fischer et al. (1979) and Ollinger (1999) was taken.

The equations for sensible and latent heat flux require the knowledge of bulk aerodynamic transfer coefficients $C_S$ and $C_L$ (eqs. 1.4 and 1.5). DYRESM assumes that the values of the bulk transfer coefficients $C_L$ and $C_S$ in the equations for latent heat and sensible heat are equal to the value of $C_M$.

### 2.5.1.2 Mean Albedo of Water Surface $\tilde{r}_{a}^{(sw)}$

Generally the albedo (reflection coefficient) of incident short wave radiation is a rather complex function of (1) the angle of the sun, (2) the relative proportion of direct and diffuse radiation, which is a function of cloud cover fraction, and (3) the surface roughness, which is a function of wind velocity and the water colour. Ollinger (1999) provides an excellent overview. DYRESM’s approach is rather simple: here the albedo $\tilde{r}_{a}^{(sw)}$ varies according to a sine function with an amplitude of 0.02, a mean $\tilde{r}_{a}^{(sw)}$ specified by the user and a phase shift such that maximum albedo occurs on the first day of the year for the northern hemisphere and on day 183 for the southern hemisphere (lowest sun angle). Antenucci & Imerito (2001) suggest a value of $\tilde{r}_{a}^{(sw)} = 0.08$ for the mean albedo.

### 2.5.1.3 Emissivity of Water Surface $\epsilon_w$

The emission of long wave radiation is computed according to eq. 1.7 on p. 14. The emissivity $\epsilon_w$ is a weak function of the surface roughness and ranges from 0.908–0.929 for wind speeds from 0–15 m/s (Gardashov et al., 1988). Imberger & Patterson (1981), however, use a value of 0.96, which is also used in this thesis.

### 2.5.1.4 Entrainment Coefficient Constant $\varepsilon$

The entrainment coefficient $E$ in DYRESM is calculated from an implicit equation that comes from the combination of equations 6.105 and 6.110 in Fischer et al. (1979)

\[
E = \varepsilon \cdot \frac{5 \tan(\phi) - 8/3E}{4E + 5C_D/\sin(\alpha)}
\]

(2.18)

\[
\varepsilon = 1/2 \cdot \eta \cdot C_k^f \cdot C_D^{3/2}
\]

(2.19)
2.5. PARAMETERS

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Table 2.21: List of numerical settings (specified in the Parameters File and Configuration File)

where $\phi$ is the longitudinal slope of the submerged river valley, $\alpha$ is the base half angle of a triangular valley cross section and $C_D$ is the drag coefficient of the river bed (see also table 2.1). The product $\eta \cdot C_D^f$ shows a great deal of variability (Fischer et al., 1979), but the estimate of $2.0 \cdot 10^{-3}$ for $\varepsilon$ by J. Imberger (pers. comm.) was used.

2.5.1.5 Buoyant Plume Entrainment Coefficient $\varepsilon_{bp}$

This coefficient is taken from Fischer et al. (1979), but does not play a role in the modelling of Lake Constance, because no bubble plume diffusers or submerged inflow pipes are present.

2.5.1.6 Shear Production Efficiency $\eta_{KE}$

Whenever two layers are mixed, of which at least one has a mean velocity greater than zero, kinetic energy is released. A certain fraction of the released kinetic energy can be utilised to reduce the required mixing energy. This fraction is determined by $\eta_{KE}$ (see also Antenucci, 2001, chap. 4). A value of 0.08 was recently determined by calibration on different lakes by Peter Yeates, CWR.

2.5.1.7 Potential Energy Mixing Efficiency $\eta_{PE}$

In case of an unstable stratification, i.e. potential energy is released when two layers are mixed, the fraction $\eta_{PE}$ of the freed energy is converted into turbulent kinetic energy. The value of 0.2 is an accepted value (priv. comm., J. Imberger).

2.5.1.8 Wind Stirring Efficiency $\eta_{ws}$

The production of turbulent kinetic energy in the uppermost layer due to the stirring action of the wind is proportional to the cube of the friction velocity $u_*$ (eq. 1.10 on p. 15) where the efficiency $\eta_{ws}$ is the constant of proportionality. The value of 0.8 was found by Peter Yeates, CWR, from a calibration on different lakes.

2.5.2 Discretization and Related Numerical Settings

Table 2.21 lists the values that determine numerical settings and discretization.

2.5.2.1 Time of Day for Output and Time Step

The model output time was chosen as 6pm and approximately coincides with the time of maximum water column stability on any day.
The model time step can no longer be freely chosen if sub-daily meteorological data is used, but the model time step must be the same as the time increment in the meteorological data (3600 s).

### 2.5.2.2 Critical Wind Speed

The mean velocity of the surface layer needs to be known in the computation of the mixed surface layer deepening. The critical wind speed $U_{\text{crit}}$ must be exceeded in order to start accelerating the uppermost layer. The process of acceleration will continue until the shear period $T_{sp}$ is reached, even if the wind speed $U$ becomes smaller than $U_{\text{crit}}$. $T_{sp}$ is defined as the minimum value of three terms:

$$T_{sp} = \min(T_i / 4, 86400 \text{ s/} \sin(\phi), 7 \text{ days})$$  \hspace{1cm} (2.20)

where $T_i$ is internal seiche period and $\phi$ is the latitude of the lake (Antenucci & Imerito, 2001). If $T_{sp}$ is reached, the shear period is recalculated, the layer velocity is reset to zero and the algorithm starts again with checking the critical wind speed criterium. A critical velocity of 3.0 m/s was used, which is exceeded about 20-30 % of the time (compare fig. 1.2).

### 2.5.2.3 Diffusion Volume Fraction and Benthic Boundary Layer Thickness

The current release of DYRESM employs two simple algorithms to simulate mixing and diffusion in the hypolimnion. Once a day, one algorithm removes the volume fraction given in the Configuration File from each layer and adds this volume (with its physical and chemical properties) to the layer above. The algorithm starts from the bottom of the lake.

The second algorithm determines for each layer below the thermocline a certain volume, removed it and dumped it into the first layer above the thermocline. The volume removed in these layers is the thickness given in the Configuration File times the difference in areas between the top and the bottom of that layer.

Both values, the diffusion volume fraction and the benthic boundary layer thickness, were set to zero, because changes to the code were made to substitute these simple algorithms with more process based ones (see sec. 2.8.1).

### 2.5.2.4 Permissible Layer Thickness

The settings for minimum and maximum permissible layer thickness ($PLT_{\text{min}}$ and $PLT_{\text{max}}$) control the allowable thickness range of layers. If a layer thickness becomes smaller than permitted, e. g. due to a withdrawal, the too thin layer is merged with the layer above or below whichever is smaller in volume. If a layer becomes too thick during a time step, e. g. after a wind event has forced layers to merge, it is split into two or more layers of equal temperature and salinity such that the thicknesses fall within the permitted range again.

From this description, it should become clear, that the value of $PLT_{\text{max}}$ strongly controls the spatial resolution of the model. This is because layers will merge and grow in thickness until they reach the upper limit, become split, and merge again, especially during turnover and in the mixed surface layer.

Furthermore some test simulations on Lake Constance and Lake Kinneret indicated that the model results are sensitive to the permitted layer thicknesses. It was shown that a thinner thickness range led to stronger erosion of the thermocline. A $PLT_{\text{max}}$ of 2–3 m proved to yield the best results, but seemed also to be a value chosen with success by many other DYRESM users.
Therefore the maximum permissible layer thickness was usually set between 2 and 3 m, while 0.5 m was assumed a reasonable minimum thickness. This also complies with the rule $PLT_{\text{max}} > 2 \cdot PLT_{\text{min}}$, otherwise the split algorithm cannot function.

### 2.5.2.5 Atmospheric Stability

For simplicity and shorter computation time, a neutral atmospheric stability was assumed, because the error in the estimation of some interpolated wind field from measurements at the shore stations around the lake — which are also in considerably different heights above the lake — is deemed much larger than any variability in the bulk aerodynamic transfer coefficients due to different degrees of stability.

### 2.5.3 Light Extinction Coefficient

If DYRESM is run in conjunction with CAEDYM, the light extinction coefficient is computed within CAEDYM from the background light extinction coefficient and the phytoplankton and suspended solids concentration, which both vary in time and depth. Otherwise a mean annual, depth averaged light extinction coefficient $k_{le}$ must be given by the user through the configuration file.

For Lake Constance, Ollinger (1999) suggests calculating $k_{le}$ from a background value $k_{bg} = 0.27 \text{ m}^{-1}$ and from the phytoplankton concentration $c_{phy}$ (expressed as chlorophyll a concentration in mg/m$^3$)

$$k_{le} = k_{bg} + k_{phy} \cdot c_{phy}$$

where $k_{phy}$ is given as 0.015 m$^2$/mg Chl$a$. Mean monthly chlorophyll concentrations in the upper 10 m vary from about 2 mg/m$^3$ in winter to 4–7 mg/m$^3$ in summer and autumn. May has the highest value with about 11 mg/m$^3$ (IGKB, 1998). As a consequence, values of $k_{le}$ range from 0.30 m$^{-1}$ in winter to 0.44 m$^{-1}$ in May. The mean phytoplankton concentration from April to November would result in $k_{le} = 0.35 \text{ m}^{-1}$.

Alternatively, it may be estimated from secchi-disk depth $d_{sd}$ measurements

$$k_{le} = C_{sd}/d_{sd}$$

where $C_{sd}$ is a constant with a value of 1.44 to 1.8 (Holmes, 1975; Chapra, 1997).

Since phytoplankton field data were lacking, but 391 secchi-disk measurements were available from the Internationale Gewässerschutzkommission für den Bodensee for the period 1974–2001, eq. 2.22 was chosen to compute $k_{le}$.

The secchi-disk depth measurements are plotted versus time in fig. 2.29A–B. The mean secchi-disk depth from these 391 measurements is 7.6 m. A typical year starts with a clear water, then in spring the first algal bloom develops and $d_{sd}$ drops from 10–15 m down to 2–3 m. Shortly after, usually in June, heavy grazing by zooplankton decimates the phytoplankton concentration and $d_{sd}$ may increase to 10 m again ("clear water phase"). The collapse of the zooplankton due to the lack of food allows the phytoplankton to recover, but with a different taxonomic composition, which causes $d_{sd}$ to be in the range of 3-6 m until autumn (Simon et al., 1998).

The vertical distribution of heat input by short wave radiation is not only dependent on $k_{le}$, but also on the incident short wave radiation. Therefore a mean value of $\overline{k_{le}}$ over a period of $n$ subsequent days, as required for the model, was calculated as a radiation weighted arithmetic mean:

$$\overline{k_{le}} = \frac{\sum_{i=1}^{n} E_{sw}(i) \cdot k_{le}(i)}{\sum_{i=1}^{n} E_{sw}(i)}$$

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where $k_{le}(i)$ is the light extinction coefficient on day $i$, calculated from eq. 2.22 and $E_{sw}(i)$ is the short wave radiation energy on day $i$. Days without a secchi-disk measurement were computed from a cubic interpolation.

The bar diagram fig. 2.29C shows as an example yearly averages calculated with eq. 2.23. The mean value for the whole period 1974–2000 is $k_{le} = 0.28 \, \text{m}^{-1}$ with $C_{sd} = 1.44$, and $k_{le} = 0.35 \, \text{m}^{-1}$ with $C_{sd} = 1.8$.

### 2.6 Temperature and Salinity Profiles

#### 2.6.1 Data availability

The Institut für Seenforschung (ISF) of the State Institute for Environmental Protection Baden-Württemberg (LfU) measures temperature, conductivity and other physical and chemical parameters at a central location in the main basin on behalf of the Internationale Gewässerschutzkommission für den Bodensee. This location is approximately halfway between the villages of Fischbach (Germany) and Uttwil (Switzerland), where the lake is deepest. The Gauss-Krüger coordinates are: $x=3528300$ and $y=5276250$.

Only data from 1980–2000 were obtained, because earlier periods could not be deliv-
ered due to technical problems at the time of request. The frequency of sampling is about 
every 2–4 weeks. The typical sampling depths are 0, 5, 10, 15, 20, 30, 50, 100, 150, 200, 
230 and 250 m, besides 1995–1996 in which additional samples at depths of 1, 2.5 and 
7.5 m were taken. The temperature measurements are accurate within 0.1 °C and the 
 conductivity was converted into salinity with eq. 2.13 as described in sec. 2.3.1.

The Zweckverband Bodensee-Wasserversorgung (ZVBWV), a large drinking water sup-
plier, is another institution that does sampling in different depths of Lake Überlingen on 
a regular basis (priv. comm. Hans-Henning Stabel, ZVBWV). Data were not requested 
because Lake Überlingen has, at times, quite different thermal and chemical properties 
compared to the main basin. The central sampling location of ISF in the main basin was 
considered more suitable for comparison with the 1D-modelling results of DYRESM. Fu-
ture model studies with CAEDYM should strongly consider these data, as one can expect 
a wide range of analysed parameters which might be difficult to obtain elsewhere in high 
temporal and spatial resolution.

Lastly, it is known that the Limnologisches Institut of University of Constance operated 
(and may still do) a lake station with a thermistor-chain in Lake Überlingen since ca. 
1986. Additionally vertical profiles of conductivity and temperature were taken at different 
locations in Lake Überlingen and in some instances at a location in the main basin near 
Meersburg (Ollinger, 1999). For similar reasons as with the data of BWV data were not 
requested, but for further modelling studies these data should be considered as well.

2.6.2 Initial Profiles

Not only are measured profiles of temperature and salinity required to compare with model 
results, but also to initialise the model at time \( t = 0 \). This information is handed over 
to DYRESM in the Initial Profile File, which contains essentially three columns: height 
above lake bottom, corresponding temperature and salinity.

DYRESM then subdivides the lake into discrete layers such that measured heights in 
the Initial Profile File coincide with the ceilings of layers and assigns the temperature 
and salinity values to these layers. In cases where two measured points are further apart 
than the permissible maximum layer thickness, DYRESM inserts the required number of 
layers in between (their thickness will then be closest to the minimum permissible layer 
thickness) and assigns to all of them the temperature and salinity values of the measuring 
point above.

As DYRESM does not interpolate the information in the Initial Profile File, the model 
will start with an unrealistic step-like distribution of temperature and salinity given the 
coarse spatial resolution of the field data. The temperature and salinity jumps, that are at 
the positions of the measuring points, will thereby introduce an artificial stability, whereas 
the layers in between measuring points represent an unrealistic neutral stability. It was 
shown that below the mixed surface layer this step-like profile was maintained until the 
next overturn of the lake.

To prevent this, all measured profiles were interpolated in Matlab with a cubic inter-
polation method to yield a spacing close to the minimum permissible layer size (usually 
0.5 m). The cubic interpolation method smoothed the data without causing any over-
shooting.
2.7 Water Levels

The State Institute for Environmental Protection Baden-Württemberg (LfU) was able to provide a complete time series of water level readings at gauge Konstanz from 01.01.1960–17.10.2000. The zero point of this gauge is at 391.89 m a.s.l. and the long term mean water level is at 395.33 m a.s.l. (Wagner et al., 1994), which is 7 cm more than the mean calculated from this data set of 15,266 days. Figure 2.30 shows for each month the means, the interval in which 90% of all values fall and the minimum and maximum observed value.

As described in sec. 2.3.13 the water level data had to be used to correct periods of inaccurate discharge measurements. Furthermore, before 1985 the discharge of the outflowing Seerhein was calculated from water levels as described in sec. 2.4.1.

2.8 Modifications to the DYRESM code

The first basic DYRESM code was written more than 20 years ago (Imberger et al., 1978). When the code was rewritten in a more recent version of Fortran several years ago, some capabilities of DYRESM were lost or are currently being reviewed and improved. One of them is the ice-cover algorithm, which is being enhanced by Katherine Prescott at CWR. Until the implementation of it, DYRESM will not accept negative air temperatures as this could lead to water temperatures below the freezing point. For a correct modelling of the heat budget, however, it was felt that negative air temperature had to be included, considering that 11.9% of all temperature observations are below zero. The minimum temperature in the data set was as low as \(-19.5\) °C. Thus, the air temperature check was switched off and a slight adjustment of the density calculations had to be done. Frequent occurrence of water temperatures that are erroneously below the freezing point was not expected since Lake Constance froze only once in the last 40 years (i.e. in winter 1962/63). This presumption is also supported by fig. 1.6, which shows that surface temperatures drop hardly below 3 °C during winter.

Peter Yeates from CWR is working on the substitution of the simple algorithms for internal and benthic boundary layer (BBL) mixing, that are implemented in the current
release of DYRESM, with ones that are process based and therefore consider forcing and strength of stratification. His latest algorithms were used for this thesis and are described in sec. 2.8.1.

With the first simulations of Lake Constance it became clear that the model results are quite sensitive to the chosen maximum permissible layer size (see fig. 3.2). This gave rise to develop a code that enforces thin layers in the surface region, but allows thicker layers near the the bottom of the lake. This proposed algorithm is discussed in sec. 2.8.2.

2.8.1 Internal and Benthic Boundary Layer Mixing
2.8.1.1 Theoretical Background

In a stratified lake, the strong density gradient of the thermocline impedes direct propagation of turbulent kinetic energy that is generated above by wind stirring, convective cooling and shear, into the hypolimnion. However, basin wide effective vertical diffusion in the hypolimnion has been determined to be much larger than molecular diffusion. Therefore energy sources to generate turbulence and turbulent mixing must be present in the meta- and hypolimnion. It is assumed that two sources of energy are available, basin wide currents causing bottom stress and, probably much more important, internal waves. Shear instabilities and wave-wave interactions will cause turbulence in the interior while shoaling and breaking of internal waves at the lake slope will be responsible for turbulence along the lake boundaries. Decay times of internal wave fields strongly suggest that a very large fraction of the internal wave energy, which eventually came from wind forcing, is dissipated at the boundaries (Imberger, 1998).

Some concepts how actual mass flux is accomplished within and out of the BBL are outlined in Imberger (2001). For intermittent and rather small areas of turbulence along the boundary, the mixed BBL sets up a density difference with respect to the stratified interior. This will lead to horizontal intrusions.

Another process is postulated to occur when a large area is affected. Gradients within the BBL normal to the bottom are negligible due to the turbulence, but gradients along the slope are the same as in the main part of the lake as they are imposed by the interior. A shear flow will form, with cold water moving upwards along the slope, and warm water flowing downwards at the outside of the BBL. This flow is driven by the tendency of the constant density surfaces to return to the horizontal (from their tilt normal to the boundary). It is shown that for a non-linear stratification, a strong net vertical heat flux will occur.

A global parametrisation of the vertical flux must, besides the stratification strength, take into account the wind forcing, which is the main source of mechanical energy inputs and which is partly transferred into internal wave energy. Stevens & Imberger (1996) show that the generation of basin-scale internal waves is a function of the Wedderburn Number and the Lake Number. The Lake number $L_N$ is defined as the ratio of restoring to overturning moments (Imberger & Patterson, 1990)

$$L_N = \frac{(Z_V - Z_G) \cdot G \cdot \beta_{tc}}{A \cdot \tau \cdot d_V}$$

(2.24)

where the term $(Z_V - Z_G)$ is the vertical distance between the centre of volume and the centre of gravity, $G$ is the gravity force exerted on the lake, $\beta_{tc}$ is the slope of the thermocline such that the thermocline intersects the water surface at the upwind shore (upwelling
event), \( A \) is the surface area on which the mean wind stress \( \tau \) acts upon and \( d_V \) is the depth to the centre of volume.

The concept of the Lake number also implies that the order of the internal seiche amplitude is \( d_{tc}/L_N \), where \( d_{tc} \) is the depth to the thermocline (Imberger, 2002). To illustrate this, suppose as an example \( L_N = 5 \), which means the restoring moment is 5 times larger than the overturning moment. To have equal moments, which would be the mechanical equilibrium situation, the restoring moment must be reduced by factor 5. This is accomplished by an angle \( \beta_{tc} \) which is only a fifth of the angle that causes upwelling and thus the thermocline displacement at the shore is also only a fifth.

Imberger (2002) derives with experimental results from Lake Kinneret (Saggio & Imberger, 2001) that the basin wide average effective vertical diffusion coefficient \( k_{z eff} \) is inverse proportional to \( L_N \)

\[
\frac{k_{z eff}}{k_{mol}} = \frac{C_{LN}}{L_N} \quad \text{for } L_N < C_{LN}
\]

where \( k_{mol} \) is the molecular diffusivity for heat \((1.4 \cdot 10^{-7} \text{ m}^2/\text{s})\) and \( C_{LN} \) is a constant for which a value of 300 was suggested.

### 2.8.1.2 Outline of the Algorithm

With this new algorithm, all layer volumes below the pycnocline are partitioned into inner volumes and BBL volumes, where the properties (temperature, salinity) are independent of each other. Each BBL volume is the product of its thickness \( h_{BBL} \) and the difference in areas between the top and the bottom of the layer, where the latter is the horizontal projection of the bottom surface area. The thickness is not given by the user as a constant, but is calculated once a day from a balance of energy supplied by bottom stress, dissipation and change in potential energy due to de- or entrainment of water from the interior into the BBL. The routine to compute bottom stresses is already implemented in DYRESM because it is needed in CAEDYM for resuspension calculations (see chap. 4.7–4.8 in (Antenucci & Imerito, 2001).

The vertical transport in DYRESM is modelled in two steps. The first step is to determine the volumes that are removed once a day from a layer and mixed into the layer above, whereby the code starts at the lowest layer and proceeds through all layers below the pycnocline upwards. The volume \( V_{mix} \) that is removed from layer \( i \) and added to layer \( i+1 \) is calculated in accordance with eq. 2.25 as

\[
V_{mix} = \frac{C_{LN}}{L_N} \cdot \frac{V_i \cdot k_{mol}}{(h_i + h_{i+1})^2} \cdot \Delta t \cdot \frac{(\Delta \rho/\Delta z)_{\text{local}}}{(\Delta \rho/\Delta z)_{\text{max}}}
\]

where \( h_i \) and \( h_{i+1} \) are the layer thicknesses, \( V_i \) is the volume of layer \( i \) and \( \Delta t \) is the time step. The second step is to split each volume \( V_{mix} \) into a volume \( \psi \cdot V_{mix} \) that is transferred between the interior volumes and a remaining volume \((1 - \psi) \cdot V_{mix} \) that is transferred between the BBL volumes:

\[
V_{mix} = V_{mix,\text{int}} + V_{mix,BBL} = \psi \cdot V_{mix} + (1 - \psi) \cdot V_{mix}
\]

\[
\psi = \frac{\tanh(B_N) \cdot (L_N - 1)}{L_N} \quad \text{for } L_N > 1
\]

\[
B_N = \frac{v_i}{L \cdot \omega}
\]
where $B_N$ is the Burger number, $v_i$ is the phase speed of an internal wave, $\omega = 2 \cdot \Omega \cdot \sin(\phi)$ is the inertial frequency at the latitude $\phi$, $\Omega$ is the angular velocity of the earth and $L$ is the representative width or length of the lake at the depth of the internal wave.

The flux from the uppermost BBL layer, which is located at the pycnocline, is transferred into the the layer above, where no partitioning into interior and BBL exists.

In cases where the flux within the BBL volumes cannot be sustained because of too small BBL volumes, the volume transfer is accomplished in the interior volumes instead. This assures the total volume flux predicted by eq. 2.26.

The Burger number is the ratio of the time it takes an internal wave to travel across the lake to the time it takes for the lake to rotate about its axis. Large Burger numbers ($\gg 1$) indicate a negligible influence of the Coriolis force and the internal oscillations will be simple gravitational seiches. Small Burger numbers ($< 1$) indicate that the system is strongly influenced by rotation and it is expected that boundary mixing becomes more, and interior mixing less significant.

The Burger number at Lake Constance will be much smaller than unity from spring to autumn. For example, with an inertial frequency of $\omega = 1.07 \cdot 10^{-4}$ rad/s, typical phase speeds $v_i$ ranging from 0.15 to 0.5 m/s (Bauerle et al., 1998) and with a conservative estimate of the length scale $L$ as 10 km, the Burger number will range from 0.14 to 0.46 and will be even smaller if $L$ is assumed to be the root of the area at the thermocline depth $(L \approx \sqrt{400 \cdot 10^6 \text{m}^2} = 20 \text{ km}$ or as the length of the main basin $L \approx 40 \text{ km}$.

### 2.8.2 Proposed Re-gridding Algorithm

With virtually every numerical model a compromise between high resolution, more accurate results and long computational time on the one hand and low resolution, less accurate model results and short computational time must be found. In contrast to a Eulerian numerical scheme with a fixed spatial grid, where differential equations are approximated by finite differences, the spatial discretisation in a Langrangian scheme is not directly related to the induced numerical error. Nevertheless a minimum spatial resolution is required as well in order to simulate the physical processes. It can be assumed that the resolution of the model must be smaller than or equal to the scale of the processes that are simulated.

One very important process in the surface layer dynamics is the heating by short wave radiation (SWR). This heat source directly stabilises the water column according to its attenuation profile. It is most of the time the largest source of heat, because the net long wave and the latent heat fluxes are almost always negative (heat loss), and the magnitude of the convective heat flux at times when it is positive is usually smaller than the SWR flux (see for example fig. 6 in Bauerle et al. (1998) and fig. 2.31). The attenuation of the SWR is a function of the light extinction coefficient $k_{le}$ only, where the spatial scale is given by the inverse of $k_{le}$. The depth $d_{rg}$, at which the flux at the water surface has been attenuated by a factor $e^{-k_{rs}}$, is $k_{rg}/k_{le}$.

The proposed re-gridding scheme makes use of this scale. At each time step, before the subroutines for heating and mixing are called, the following steps are executed:

1. A depth $d_{rg} = k_{rg}/k_{le}$ is determined to which re-gridding will approximately occur. For the time being, $k_{rg}$ is set to unity.

2. A distance $L_{rg} = d_{rg}/n_{rg}$ is calculated, where $n_{rg}$ is a number that is currently set to a value of $5 \cdot k_{rg}$.

---

5The re-gridding algorithm has evolved from ideas by J. Antenucci, D. Horn, J. Imberger, A. Imerito, K. Prescott, P. Yeates and myself. Coding was done by A. Imerito.
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3. Then, starting from the water surface the code proceeds downwards in steps of \( L_{rg} \) to insert new layer boundaries into the existing layer structure until the depth \( d_{rg} \) is reached. If a new layer boundary is too close (\(< L_{min} \)) to an existing one, the insertion will be skipped, and the existing layer boundary will be the new starting point for the next step downwards of length \( L_{rg} \). The value of \( L_{min} \) is currently set to 10 cm.

This procedure ensures that — independent of the existing layers — there will be at least \( n_{rg} \) numbers of layers between the water surface and a depth \( d_{rg} \). Thus the user can specify a reasonably large maximum permissible layer thickness (e.g. 10 m instead of 1–3 m) and will still have a well resolved surface layer region.

The preliminary subroutine was tested with the web-release code of DYRESM on Lake Constance and appeared to meet the expectations. However, the following points need still to be discussed:

1. The value of \( k_{rg} \). A value of 3 is suggested for \( k_{rg} \) so that 95 % of the SWR has been attenuated at \( d_{rg} \).

2. The number of layers that are inserted per length \( 1/k_{le} \) (i.e. currently 5) is arbitrary.

3. The criterion \( L_{min} \), which prevents the creation of too thin layers. At the moment it has an absolute value of 10 cm, but it is suggested that \( L_{min} \) is a function of \( 1/k_{le} \), e.g. \( L_{min} = 0.2 \times 1/k_{le} \).

4. New layer boundaries might be inserted exponentially spaced instead of equally spaced. Exponential spacing would lead to layers that receive equal amounts of SWR energy.
Chapter 3

Results and Discussion

The first section of this chapter illustrates, with an example, that the DYRESM code without a diffusion algorithm is not capable of reproducing the apparently enhanced mixing occurring in the metalimnion of Lake Constance. Furthermore, it is shown that the model is sensitive to the choice of the maximum permissible layer thickness.

The second section presents the results of a calibration of three parameters. The diffusion algorithm introduced a constant that controls the amount of diffusion given a certain Lake Number. This constant appears to vary from lake to lake and a value specific for Lake Constance had to be found. In addition to this parameter, the outcome of the first section indicated a need to consider the maximum permissible layer thickness in the calibration procedure. Furthermore, it is known that the wind speed measured at the shore stations is not representative for the average wind speed over the lake surface. A factor that multiplies the wind speed of the shore stations was the third parameter included in the calibration.

To limit the number of simulation runs for the calibration, two assumptions were made. One was that the value of the light extinction coefficient is not subject to a calibration but given by field measurements within a comparatively small range or by a water-quality model. The other was that the model has a low sensitivity to the minimum permissible layer thickness. These assumptions are discussed in the third and fourth section of this chapter.

The fifth section attempts to verify the calibration results with time periods that were not used for the calibration. It makes clear that the quality of the prediction is very different from year to year, but also that field data of much higher sampling frequency is necessary to exclude errors caused by internal wave displacements.

The sixth section investigates the model sensitivity to the meteorological variables air temperature and cloud cover fraction. Additionally, two simulations with modified meteorological variables are compared to the unmodified cases.

In the following comparisons of simulations with field data, it must be kept in mind that some of the larger errors, particularly in the metalimnion during autumn, may be due to displacements of isotherms by internal waves. This is a three-dimensional phenomenon, superimposed on the mean one-dimensional temperature stratification, which of course cannot be predicted by DYRESM. Figures 16 and 17 in Bauerle et al. (1998) can be consulted to quantify the potential temperature deviation from the mean structure. It is no less than 7 °C difference between a measurement that took place in 10 m depth in...
the morning of day 232 of year 1992 when the trough of the internal wave passed, and a
measurement at the same depth in the evening of day 233, when the crest passed. The
internal wave had an amplitude of about 5 m. The coarse temporal resolution of the field
data (2–4 weeks) and the lack of other stations make it impossible to filter the data to
remove the internal wave signal.

All analyses in this chapter are based on the state variable temperature. The other
state variable, salinity, is not considered at all. This is because the focus is on the strat-
ification period (spring to late autumn) during which the contribution of salinity to the
density stratification is negligible (sec. 1.2.4.1).

Temperature contour plots are printed as gray scale plots in the main text and are
repeated as colour plots in the appendix A. Cross references are given in the figure
captions.

3.1 Necessity of a Diffusion Algorithm

Simulation results are unsatisfactory if the standard code of DYRESM is applied to Lake
Constance, which does not include the diffusion algorithm discussed in sec. 2.8.1. This is
shown by fig. 3.1 (gray scale) and fig. A.1 (colour), in which temperature predictions are
compared with field data for the 3-year period 1980–1982. Temperature errors range from
about -5 to +4 °C. A general observation is that the temperatures in the mixed surface
layers are overestimated in the first half of the stratification period and that temperatures
are considerably underestimated in the depth range 10–30 m during the second half of the
stratification period.

One reason for the deviations might be as follows: If the stratification commences too
early in the model (most pronounced in 1980), heat is retained in the surface layers, while
in reality the weaker stratification still allows mixing of heat at the surface down to greater
depths. An earlier beginning in the model than in reality might be the consequence of a
wind event, that led to the destruction of a first, but still weak, stratification in reality
but not in the model. In reality surface layer deepening is a continuous process, but in the
model it must occur in discrete steps. It could be imagined that the accumulated wind
energy might just fail to provide the required energy to mix a complete layer and deepening
stops. A too early start of stratification could also come from measured meteorological
forcing that is not representative for the lake for a few days. A simple scenario could be
fog that is spread over the lake surface for half a day, whereas the meteorological stations,
that are up to 45 m above the lake, receive sunshine beginning in the early morning. In
both cases the deviation between model and reality is then accelerated because there is a
positive feedback such that retaining heat in the surface layers increases stability of the
stratification which in turn prevents mixing of heat to deeper layers. As a consequence,
the surface layers can become much too warm in the model for a few weeks in late spring
or early summer and at the same time an initial heat deficit in the depth region between
15–30 m develops.

However, field data also show that there is hardly any distinct seasonal thermocline.
Instead, temperatures vary smoothly over a large vertical extent. Hypolimnetic tempera-
tures are not reached until 40–50 m depth. This is in contrast to the model results
that predict a strong seasonal thermocline at a depth of about 15 m (in 1980) such that
hypolimnetic temperatures are already reached in about 20 m depth (see also fig. 3.2).
Consequently, temperatures are too cold below the modelled thermocline, explaining the
large negative errors shown in the bottom panel of fig. 3.1 below 15 m depth.

That DYRESM correctly simulates the most important physical processes in lakes has
3.1. NECESSITY OF A DIFFUSION ALGORITHM

Figure 3.1: Temperature contour plots of a simulation without the diffusion algorithm (top), IGKB field data (middle) and the temperature differences between simulation and field data (bottom). Dots indicate measurements. The wind multiplication factor (WMF) is 1.3, the permissible layer thickness (PLT) is 0.5–2.5 m, light extinction coefficient \( k_{le} \) is 0.35 m\(^{-1}\) and the model was started on 15-Jan-1980. (Contouring grid: \( dt = 2 \) days for simulation, \( dt = 4 \) days for field data and differences, \( dz = 2 \) m, linear interpolation.). For a colour version see fig. A.1
Common parameters: \( \text{WMF} = 1.3; \ C_{LN} = 1; \ \text{plt}_{\text{min}} = 0.5 \text{ m}; \ k_t = 0.35 \text{ m}^{-1} \).

![Temperature profiles of simulations (lines) and IGKB field data (squares) on three different days in 1980. The only difference between simulations is the maximum PLT. Common parameters: WMF= 1.3, PLT_{\text{min}} = 0.5 \text{ m} and \ k_t = 0.35 \text{ m}^{-1}. Model start was 15-Jan-1980. The profiles shown by the thick black line correspond to the simulation of fig. 3.1.

been proven by successful applications to other lakes. Therefore, the observation of a smeared thermocline compared to the sharp simulated thermocline, might be another hint for unusually strong turbulent diffusion occurring in Lake Constance compared to other lakes (see also sec. 1.2.4.4 on p. 9). This process, however, cannot be modelled with the standard version of DYRESM, but requires an algorithm for diffusion in the meta- and hypolimnion.

A secondary result from testing the standard code of DYRESM was that simulations can differ substantially dependent on the imposed maximum permissible layer thickness (\( \text{plt}_{\text{max}} \)), even if only reasonable values are chosen. This is illustrated by temperature profiles of three simulations that differ in \( \text{plt}_{\text{max}} \) only (fig. 3.2). It can be concluded from these profiles that the smaller the layer thickness, the sharper the thermocline. The choice of the minimum permissible layer thickness (\( \text{plt}_{\text{min}} \)) had insignificant effects.

3.2 Calibration

With the implementation of the internal and benthic boundary layer mixing algorithm to simulate turbulent diffusion in the hypo- and metalimnion (or short “Lake Number mixing”), one more non-generic parameter, \( C_{LN} \), was introduced to DYRESM. It is the constant of proportionality in eq. 2.25, which expresses that the effective vertical diffusion is inversely proportional to the Lake Number. \( C_{LN} \) determines thus the intensity of diffusion given a certain state of internal wave activity after perturbation by wind, which is represented by the Lake Number. The value of \( C_{LN} \) is probably a strong function of the lake geometry, e. g. the bottom slope at metalimnion depth and the deviation from a regularly shaped basin. The bottom slope determines whether internal waves brake or reflect. Until future research allows parameterisation of these factors, \( C_{LN} \) remains a calibration value.
Ollinger (1999) uses a wind multiplication factor (WMF) of 1.31 to convert values of wind speed at station Konstanz to values representative for Lake Überlingen. Zenger et al. (1990) point out that strong winds are more frequently observed at the northern shore of the main basin than at Unteruhldingen at the shore of Lake Überlingen. They also estimate that wind speed at the main basin may be on average 20% larger than at Lake Überlingen, which would increase the wind multiplication factor from 1.31 to as much as 1.57 for the conversion from station Konstanz to the main basin. On the other hand, the prepared wind speed time series is already an average of the stations Konstanz, Gütingen and Friedrichshafen (sec. 2.2.5), so that the actual factor may be close to 1.3.

Generally, measured forcing data should not be calibrated unless the accuracy of the measurements is low or unless it is doubtful that the measurements are representative. The large conversion factors found in the literature tell that the latter is true. Additionally, wind speed is probably one of the most important input variables in the model because wind can be expected to supply most of the mechanical energy that is required for mixed layer deepening and turbulent diffusion through internal wave activity. High sensitivity can be anticipated from the fact alone that the acting wind force is proportional to the squared wind speed. Therefore, it seemed necessary to consider the wind multiplication factor as the third calibration parameter.

The calibration of diffusion intensity $C_{LN}$, maximum permissible layer thickness $PLT_{max}$ and wind multiplication factor WMF is a three-dimensional optimisation problem. For simplicity it was reduced into 2 two-dimensional problems: At first, an optimal parameter set for $C_{LN}$ and $PLT_{max}$ was determined while WMF was assumed constant at a value of 1.3. If 1.31 and 1.57 are suggested for the conversion from station Konstanz to Lake Überlingen and to the main basin, respectively, and if the wind speed time series is already an average over several shore stations, a value of 1.3 can be considered to be close to the actual value for the whole Upper Lake Constance. The objective of the first step was to fix the value of $PLT_{max}$ and to get a first approximation of $C_{LN}$ (sec. 3.2.1).

In the second step, the value of $PLT_{max}$ was held constant at the value found in the first step, while $C_{LN}$ and WMF were calibrated (sec. 3.2.2). This calibration in two steps concentrates on the two parameters that have a physical meaning, diffusion intensity and wind speed, and less on the maximum layer thickness, for which the influence on the model behaviour is much more difficult to understand and predict.

### 3.2.1 Calibration of Diffusion Intensity and Maximum Layer Thickness

The calibration was performed with the 3-year periods 1980–1982 and 1995–1997. The first period was chosen as representative for the time where the lake was still at its highest trophic level (phosphorus concentration during turnover about 70-80 mg/m$^3$), whereas the second was deemed to represent a low trophic level (15-20 mg/m$^3$). The first period could not be shifted a few years backwards where phosphorus and algal concentrations were even higher, because neither short wave radiation measurements nor field data were received for the time before 1980. The second period includes the years 1995 and 1996, in which sampling frequency and depth resolution were higher than in the other years (sec. 2.6.1).

A total number of 97 simulations were run and only the maximum permissible layer thickness ($PLT_{max}$) and $C_{LN}$ were varied. $PLT_{max}$ values ranged from 2.0 to 5.0 m and $C_{LN}$ values from 1 to 7200. The most likely value for the light extinction coefficient $k_{le}$ was taken for the calibration. A value of 0.35 m$^{-1}$ is supported by eq. 2.21 as an average for the stratified period as well as from eq. 2.23 as an average in each of the two 3-year periods if $C_{sd} = 1.8$ is taken. A value of 0.5 m seemed to be a reasonable assumption for $PLT_{min}$. The low sensitivity of the model to this parameter is shown in sec. 3.4.

As an objective measure of the model performance, three numbers were calculated from
Model performance dependent on maximum permissible layer thickness ($PLT_{max}$) and intensity of "Lake Number Mixing" $C_{LN}$.

Period: 19800115−19830111, number of values per simulation: 222, depth range 0−35 m, WMF=1.3, $PLT_{min}$=0.5 m, $k_{le}=0.35$ 1/m.

Figure 3.3: Contours of model performance dependent on parameters max. perm. layer thickness $PLT_{max}$ and Lake Number mixing intensity $C_{LN}$. The performance is expressed with three different measures: arithmetic mean of temperature differences (=errors) between model and field data, standard deviation of errors and root mean squared value of errors (eqs. 3.2−3.4). Only the depths between 0 and 35 m were considered. Crosses correspond to successful simulations (46), squares to aborted simulations due to an error (1). All simulations ran from 15-Jan-1980 to 11-Jan-1983. Other parameters: WMF=1.3, $PLT_{min}$=0.5 m, $k_{le}=0.35$ m$^{-1}$.

Model performance dependent on maximum permissible layer thickness ($PLT_{max}$) and intensity of "Lake Number Mixing" $C_{LN}$.

Period: 19950502−19980707, number of values per simulation: 494, depth range 0−35 m, WMF=1.3, $PLT_{min}$=0.5 m, $k_{le}=0.35$ 1/m.

Figure 3.4: As fig. 3.3, but 49 successful simulations and 1 aborted simulation from 21-Feb-1995 to 7-Jan-1998.

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each simulation: the arithmetic mean of temperature errors $\tau$, the standard deviation of the errors $\sigma_e$ and the root mean squared value of the errors $rms_e$ (eqs. 3.2–3.4).

$$
\epsilon(t_i, d_j) = \theta_{\text{sim}}(t_i, d_j) - \theta(t_i, d_j)
$$

$$
\tau = \frac{1}{\sum_{i=1}^{n} m_i} \cdot \sum_{i=1}^{n} \sum_{j=1}^{m_i} \epsilon(t_i, d_j)
$$

$$
\sigma_e = \left( \frac{1}{\sum_{i=1}^{n} m_i - 1} \cdot \sum_{i=1}^{n} \sum_{j=1}^{m_i} (\epsilon(t_i, d_j) - \bar{\epsilon})^2 \right)^{1/2}
$$

$$
rms_e = \left( \frac{1}{\sum_{i=1}^{n} m_i} \cdot \sum_{i=1}^{n} \sum_{j=1}^{m_i} \epsilon(t_i, d_j)^2 \right)^{1/2}
$$

A temperature error at field measurement depth $d_j$ on day $t_i$ is the difference of the simulated temperature $\theta_{\text{sim}}$ at that depth at 6 pm and the field measurement value $\theta$. $m_i$ is the number of different sampling depths on a sampling day $i$ and $n$ is the number of days on which measurements were taken. Only sampling depths shallower than 35 m were considered in the calibration, so that $m = 6$ in 1980–82 and $m = 6 \ldots 9$ in 1995–97. There are 37 days in the period 1980–82 and 68 days in the period 1995–97 for which field data were collected. Each simulation statistic is based on 222 (1980–82) and 494 (1995–97) single error values. The three statistical numbers are contoured as functions of $PLT_{\text{max}}$ and $C_{LN}$ in figs. 3.3 and 3.4.

The mean value suggests combinations that are on a line from about $PLT_{\text{max}} = 5$ m and $C_{LN} = 100$ to $PLT_{\text{max}} = 2$ m and $C_{LN} = 800 \ldots 2000$. The mean value, however, must always be analysed in conjunction with the standard deviation because a desired mean of zero can be achieved by enormous, but symmetrically distributed errors. Combinations that have a mean of zero and at the same time a small standard deviation reduce the line of possible combinations to layer thicknesses from 2.0–3.0 m and $C_{LN} = 2000$ ($\sigma_e \approx 1.1 \degree C$) for 1980–1982 and to a single point at 2.0 m and $C_{LN} = 1000$ ($\sigma_e \approx 1.3 \degree C$) for 1995–97.

The rms value suggests that the optimal parameter combination lies in a band with one end at about $PLT_{\text{max}} = 3 \ldots 4$ m and $C_{LN} = 1000$ and the other end at the bottom right corner of the panels with $PLT_{\text{max}} = 2$ m and $C_{LN} = 7200$. A minimum rms value means that this is the most reliable simulation in the sense that the sum of the errors, regardless of their sign, is smallest. Due to the summation of squares, large errors are weighted more than small errors. The parameter combinations from the analysis of the rms value do not coincide with the ones that are considered optimal in terms of mean and standard deviation: the curve of minimum rms values for layer thicknesses between 2.0–3.0 m runs roughly along the $+0.3$ and $+0.4 \degree C$ isolines of the mean error 1980–82 and 1995–97 respectively.

Figures 3.5 and 3.6 try to illustrate how temperature profiles of simulations with optimal rms value compare to profiles of simulations with a mean of zero at low standard deviation. The differences between these simulations are generally small. The simulation with $C_{LN} = 3000$ and $PLT_{\text{max}} = 2.5$ m achieves by far the best fit in 30 m depth, where the others perform relatively poorly. Although this parameter combination tends to overestimate temperatures at 15 and 20 m depth, it was considered as the best combination. Therefore, the second step of the calibration used a fixed value of $PLT_{\text{max}} = 2.5$ m, and a value of 3000 served as the starting point for the calibration of $C_{LN}$.

This calibration also confirms that model results improve with increasing layer thickness if $C_{LN}$ is smaller than about 1000. This was shown in the temperature profiles of fig. 3.2, where $C_{LN} = 1$. The opposite becomes true if $C_{LN}$ is larger than about 3000. Layer thickness seems to play a minor role for $C_{LN}$ in between 1000 and 3000.
CHAPTER 3. RESULTS AND DISCUSSION

08−Jul−1980
(1980190)

Common parameters: wmf = 1.3; plt\_min = 0.5 m; k\_le = 0.35 m\(^{-1}\).

09−Sep−1980
(1980253)

Common parameters: wmf = 1.3; plt\_min = 0.5 m; k\_le = 0.35 m\(^{-1}\). Model start was 15-Jan-1980.

07−Oct−1980
(1980281)

Common parameters: wmf = 1.3; plt\_min = 0.5 m; k\_le = 0.35 m\(^{-1}\).

01−Jul−1997
(1997182)

Common parameters: wmf = 1.3; plt\_min = 0.5 m; k\_le = 0.35 m\(^{-1}\).

19−Aug−1997
(1997231)

Common parameters: wmf = 1.3; plt\_min = 0.5 m; k\_le = 0.35 m\(^{-1}\).

07−Oct−1997
(1997280)

Common parameters: wmf = 1.3; plt\_min = 0.5 m; k\_le = 0.35 m\(^{-1}\).

Figure 3.5: Temperature profiles of simulations (lines) and IGKB field data (squares) on three different days in 1980. Two simulations correspond to an optimal combination of C\(_{LN}\) and PLT\(_{max}\) in terms of a minimum rms value, one simulation in terms of a mean of zero at low standard deviation. Common parameters: WMF=1.3, PLT\(_{min}\) = 0.5 m and k\(_{le}\) = 0.35 m\(^{-1}\). Model start was 15-Jan-1980.

Figure 3.6: As fig. 3.5, but profiles from 1997 with model start on 21-Feb-1995.
3.2.2 Calibration of Diffusion Intensity and Wind Multiplication Factor

To calibrate diffusion intensity in conjunction with the wind multiplication factor, the same time periods, the same minimum permissible layer thickness and the same light extinction coefficient as in the previous section were used. In 84 simulations $C_{LN}$ was varied from 600 to 4800 and WMF from 1.0 to 1.6. The model performance is plotted in figs. 3.7 and 3.8.

Most isolines have a diagonal direction, which means that $C_{LN}$ must be reduced to achieve the same performance if WMF is increased. This behaviour can be explained. A certain model performance, which is in this analysis to a large part determined by the correct prediction of the thermocline structure, stays constant if the diffusive flux remains roughly constant. The diffusive flux is a function of the Lake Number and $C_{LN}$. As the Lake Number is inversely proportional to the wind speed squared, the diffusive flux remains constant if $C_{LN} \cdot (WMF \cdot U)^2 = const$. For example, if WMF is changed from 1.3 to 1.5 (factor of 1.15), $C_{LN}$ must be decreased by a factor $(1/1.15)^2 = 0.75$ to keep the diffusive fluxes constant.

Figures 3.7 and 3.8 show that a WMF ranging from 1.4 to 1.5 and a diffusion intensity $C_{LN}$ ranging from 1000 to 2000 yield the best results in terms of all three criteria (mean error, standard deviation and rms value). Compared to the parameter combination of the first calibration step (1.3, 3000), the mean improves by about 0.1-0.2 °C, the standard deviation remains approximately constant and the rms value decreases by about 0.075 °C.

A parameter combination of $C_{LN} = 1500$ and WMF= 1.45 is suggested as optimum because higher values of WMF lead to worse results in the mean error without improving the standard deviation or the rms value. This new optimal combination has a diffusive flux that is 38 % less than the one from the first step of the calibration. The reason is that the higher wind speeds lead to deeper mixing and later onset of stratification. The increase in initial heat input in depth equal to or greater than 30 m reduces the necessity of heat transport by diffusion during stratification. This is shown in fig. 3.9 with three temperature profiles for spring 1997. It appears also that the prediction of the depth and temperature of the mixed surface layer improves slightly. This is illustrated with temperature profiles from autumn 1996 (fig. 3.10).

The calibration result for WMF of about 1.45 agrees with the values given in the literature (1.3 and greater). This also indicates that DYRESM is capable of excellent predictions if supplied with correct forcing data.

The great improvement achieved by this calibration can be seen if figs. 3.1 and 3.11 or colour figs. A.1 and A.2, respectively, are compared.

3.3 Sensitivity to the Light Extinction Coefficient

In the calibration described in the previous section the values of the light extinction coefficient $k_{le}$ and the minimum permissible layer thickness $PLT_{min}$ were held constant. The value of $k_{le}$ should not result from a calibration but should be determined by measurements. This is because $k_{le}$ is not a constant in some empirical equation like $C_{LN}$ or some setting like $PLT_{min}$ inherent in a numerical model.

However, the value $k_{le}$ is similar to the wind multiplication factor relatively uncertain. One reason is that $k_{le}$ is usually not measured directly but is calculated with empirical formulae from other parameters like secchi-disk depth or phytoplankton concentrations. The main reason for the uncertainty, however, is the difficulty to find a representative depth and time averaged coefficient. If the water-quality model CAEDYM is coupled to DYRESM, $k_{le}$ is calculated within CAEDYM as a function of time and depth from phytoplankton and suspended solids. A calibration of a mean yearly coefficient for DYRESM alone would have become useless. Therefore $k_{le}$ was excluded from a calibration.
Model performance dependent on wind multiplication factor (WMF) and intensity of "Lake Number Mixing" $C_{\text{LN}}$. Period: 1980015−1983011, number of values per simulation: 222, depth range 0−35 m, PLT = 0.5−2.5 m, $k_{\text{le}} = 0.35$ 1/m

mean temperature error
standard deviation of temperature errors
rms value of temperature errors

Figure 3.7: Contours of model performance dependent on parameters wind multiplication factor WMF and Lake Number Mixing intensity $C_{\text{LN}}$. The performance is expressed with three different measures: arithmetic mean of temperature differences (=errors) between model and field data, standard deviation of errors and root mean squared value of errors (eqs. 3.2−3.4). Only the depths between 0 and 35 m were considered. Crosses correspond to successful simulations (31), squares to aborted simulations due to an error (11). All simulations ran from 15-Jan-1980 to 11-Jan-1983. Other parameters: PLT=0.5−2.5 m, $k_{\text{le}} = 0.35$ m$^{-1}$.

Figure 3.8: As fig. 3.7, but 40 successful and 2 aborted simulations from 21-Feb-1995 to 7-Jan-1998.
3.3. SENSITIVITY TO THE LIGHT EXTINCTION COEFFICIENT

Figure 3.9: Temperature profiles of simulations (lines) and IGKB field data (squares) on three different days in spring 1997. A larger wind multiplication factor improves heat input in depths of 30–50 m. Common parameters: PLT=0.5–2.5 m and $k_e = 0.35$ m$^{-1}$. Model start was 21-Feb-1995.

Figure 3.10: As fig. 3.5, but profiles from autumn 1996. A higher wind multiplication factor slightly improves prediction of mixed surface layer depth and temperature.
Figure 3.11: Temperature contour plots of a simulation with optimal parameters $PLT_{\max} = 2.5$ m, $C_{LN} = 1500$ and WMF = 1.45 from calibration (top), IGKB field data (middle) and the temperature differences between simulation and field data (bottom). Dots indicate measurements. $PLT_{\min} = 0.5$ m, $k_{le} = 0.35$ m$^{-1}$ and the model was started on 15-Jan-1980. (Contouring grid: as fig. 3.1). For a colour version see fig. A.2.
3.3. SENSITIVITY TO THE LIGHT EXTINCTION COEFFICIENT

It is therefore rather important to know if the model performance would change at all if the \( k_{le} \) were different due to the uncertain estimation of that value. For each of the 3-year periods that were used in the calibration, the optimal parameter combination \( PLT_{max} = 2.5 \text{ m}, C_{LN} = 1500 \) and \( WMF = 1.45 \) was used for two more simulations in which \( k_{le} \) was varied by \( 0.10 \text{ m}^{-1} \) (29 %).

The resulting model performance is plotted in fig. 3.12 A–B versus the light extinction coefficient. The number near each line is the normalised sensitivity coefficient \( \chi \)

\[
\chi = \frac{\Delta Y}{\Delta X} \frac{Y}{X}
\]

where \( Y \) is the model performance (e. g. the rms value) at the parameter value \( X \), which is varied by \( \Delta X \) causing a \( \Delta Y \). For example, a value of \( \chi = -0.19 \) at the rms line in fig. 3.12 A means that a 1 % increase of \( k_{le} \) improves the model performance in terms of the rms value by 0.19 % at the midpoint \( k_{le} = 0.30 \text{ m}^{-1} \). Hence, large absolute values of \( \chi \) mean high sensitivity, values close to zero low sensitivity, a value of 1 means a direct translation of a change in input to a change in output. High sensitivity should be an incentive to direct effort to a more accurate determination of the parameter value.

It can be seen that for almost all performance criteria and for both periods the simulations improve slightly with increasing light extinction coefficient. The model performance of the simulations in fig. 3.12 A is plotted versus depth in fig. 3.13. This figure suggests as a reason for the improvement, that the overestimation of temperatures in 10–20 m depth is strongly reduced if \( k_{le} \) is increased because more heat is retained in shallower depths. The overall performance becomes better even though the 30 m depth predictions become worse.

In summary, neither the lower value of \( k_{le} \) (0.25 \text{ m}^{-1}) nor the higher value (0.45 \text{ m}^{-1}) can cause the differences in the model performance that can be seen if \( PLT_{max}, C_{LN} \) or \( WMF \) are varied. This is illustrated if temperature profiles with the three different values of \( k_{le} \) in 3.14 are compared with figs. 3.2 and 3.5.
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Figure 3.13: Mean temperature error, standard deviation and rms error (eqs. 3.2–3.4) versus depth for different light extinction coefficients. Model period 1980–82. These three simulations correspond to the three simulations in fig. 3.12 A.

Figure 3.14: Temperature profiles of simulations that only differ in the light extinction coefficient (lines) and IGKB field data (squares) on three different days in 1980. PLT=0.5–2.5 m, $C_{LN}=1500$ and WMF= 1.45.
3.4 Sensitivity to the Minimum Permissible Layer Thickness

PLT\textsubscript{min} and PLT\textsubscript{max} are parameters that can only be determined by a calibration. A parameter may not be part of calibration if the model is not sensitive to a change of its value. Every reduction in the number of calibration parameters greatly reduces the number of required simulation runs. A few runs with random parameter combinations indicated that a change PLT\textsubscript{min} causes only insignificant changes in the model result. PLT\textsubscript{min} was therefore not included in the calibration. That the model is insensitive to PLT\textsubscript{min} at the PLT\textsubscript{max}, C_{LN} and WMF parameter combination that resulted from the calibration procedure is shown in fig. 3.12 C–D. Almost all sensitivity coefficients have absolute value close to or equal to zero. Therefore, the exclusion of this parameter from the calibration was valid.

3.5 Verification

A calibrated model should always be tested on a data set which was not used in the calibration. This is called model verification (Schnoor, 1996). This was done with two further simulations at time periods different from the calibration periods. One simulation was from 14-Feb-1984 to 20-Dec-1991 (almost eight years) and the other from 07-Jan-1998 to 30-Dec-2000 (almost three years). It was intended to run the first one until 21-Feb-1995, which was the first day of the second calibration period, but an error prevented continuation.

The criteria used are the same as in the calibration: mean error, standard deviation and rms value (eqs. 3.2–3.4). For each of the two simulations these three statistical values were calculated for one year periods from 01-Jan to 31-Dec. Only the first and last years of each simulation were a bit shorter. The result of this analysis is plotted in fig. 3.15. It is remarkable how variable the model performance is. The model prediction is particularly bad in the years 1984–1986. The only year where all three criteria are better than or equal to the calibration periods is 2000. The vertical distribution of the model performance is plotted in fig. 3.16 for the best year (2000) and the worst year (1986). Both simulations have the largest errors for the depths 10, 15 and 20 m, where temperatures are always overestimated. The worse performance of 1986 is not a deviation from this general pattern,
CHAPTER 3. RESULTS AND DISCUSSION

Figure 3.16: Mean temperature error, standard deviation and rms error (eqs. 3.2-3.4) versus depth for the best (2000) and the worst year (1986) of the verification periods.

Figure 3.17: Temperature profiles of a simulation (line) and IGKB field data (squares) on three consecutive sampling days including the profile of the largest deviation of 1984 on 11 September. WMF= 1.45, $C_{LN} = 1500$, PLT=0.5–2.5 m, $k_{le} = 0.35$ m$^{-1}$.
3.5. VERIFICATION

Figure 3.18: Temperature contour plots of the period of worst model performance (top), IGKB field data (middle) and the temperature differences between simulation and field data (bottom). Dots indicate measurements. WMF=1.45, $C_{LN}=1500$, PLT=0.5–2.5 m, $k_{le}=0.35$ m$^{-1}$, model start 14-Feb-1984. (Contouring grid: as fig. 3.1). For a colour version see fig. A.3
but only an extreme case. The light extinction was according to fig. 2.29 on p. 54 not significantly different from the average of the period 1974–2000.

The temperature contour plot (fig. 3.18) of the years 1984–86 reveals that a large fraction of the statistical error of each year seems to be caused by only a few profiles with extreme deviations in 10, 15 or 20 m depth. In 1984 it is on 11 September, in 1985 on 11 June, and in 1986 on 13 May and three consecutive profiles on 8 July, 5 August and 3 September.

The temperature profiles of 11-Sep-1984 and the first sampling days before and afterwards are plotted in fig. 3.17. On 7 August the typical overestimation of temperatures at 10–20 m depth is obvious, which becomes much stronger on the next sampling day on 11 September. Astonishingly, one sampling day later, on 16 October, field data and simulation match almost perfectly again. A closer look on the field data of 11 September reveals that temperatures at 15, 20 and 30 m depth drop by about 1–2 °C in comparison to the previous sampling profile, whereas at 10 m depth temperature rises by 2–3 °C due to deepening of the mixed surface layer. As the deepening does not reach the 15 m depth, the only remaining processes that can cause a temperature change below are diffusion and absorption of short wave radiation energy. However, both processes would cause an increase in temperature and not the observed decrease. Thus it is suspected that a displacement of isotherms by an internal wave is responsible for the extreme temperature deviation between model and field data in the metalimnion on 11 September. In the introduction of the chap. 3 it was mentioned that internal waves can cause temperature deviations from the mean structure of several degrees. It can be presumed that some of the other extreme errors in thermocline depths in other years have the same reason.

3.6 Results from Changes in Meteorological Forcing

3.6.1 Sensitivity to Air Temperature and Cloud Cover Fraction

It is not only interesting to know the model sensitivity to parameters like the light extinction coefficient, but also to forcing data. The higher the sensitivity the more important is a forcing variable. High sensitivity means that inaccurate measurements or measurements with high standard deviations directly effect the model result and vice versa.

According to fig. 3.19 this is clearly the case with the two variables air temperature $\theta_{air}$ and cloud cover fraction (CCF). These two were chosen as examples out of the six meteorological input variables. The influence of the wind speed can be deduced from figs. 3.7 and 3.8, and the mass and heat fluxes by precipitation are negligible. The influence of short wave radiation is included in the sensitivity to CCF, because a change in CCF causes a change in the short wave radiation ratio according to fig. 2.7 as well as a very small change in incoming long wave radiation (eq. 1.8).

This sensitivity analysis was done on the same time periods as the other sensitivity analysis and the calibrations (i. e. 1980–82 and 1995–97). The air temperature time series was varied by adding or subtracting 2 °C to every air temperature measurement. This changed the mean air temperature from 9.3 °C to 11.3 °C and 7.3 °C respectively. Cloud cover fraction was varied by a fraction of ±0.1 changing the mean from 0.70 to 0.60 and 0.80 respectively.

A minimum rms value and a mean error closest to zero for the case of unmodified air temperatures could mean that the air temperature time series prepared from shore stations is representative for the lake. It is also impressive that a 2 °C change in air temperature causes an almost 1 °C change in mean water temperature in the upper 35 m (fig. 3.19).

An increase in cloud cover fraction would improve the model performance with respect to all three criteria and in both investigated time periods. Cloud cover fraction can only cause a change of 17 % in long wave radiation between a sky with no clouds and a sky
3.6. RESULTS FROM CHANGES IN METEOROLOGICAL FORCING

Figure 3.19: Mean temperature error, standard deviation and rms error (eqs. 3.2-3.4) as functions of the mean air temperature $\theta_{\text{air}}$ (A and B) and the cloud cover fraction CCF (C and D) for two 3-year periods. The numbers adjacent to the lines are the normalised sensitivity coefficients (eq. 3.5). WMF=1.45, PLT=0.5-2.5 m, $C_{Laste}=1500$, $k_{le}=0.35$ m$^{-1}$. In 1995-97 two simulations could not be included due to an error.

completely covered by clouds (eq. 1.8). Thus the improvement must be due to the reduced short wave radiation. Relatively frequent fog close to the ground during periods of a stably stratified atmosphere but short wave measurements on shore at an elevated level may lead to an overestimation of incident short wave radiation at water surface level. The relatively high sensitivity to cloud cover fraction is of particular concern because short wave radiation measurements at Lake Constance are not available at all before October 1977 and were only delivered for the period 1980–80. Short wave radiation for the time period before 1980 was therefore calculated from cloud cover fraction by a regression analysis as described in sec. 2.2.1. These estimated data may not be of sufficient quality because the scatter in the relationship between short wave radiation ratio and cloud cover fraction was high (fig. 2.7) and furthermore the observation frequency of cloud cover fraction was lower before 1980 than after 1980 (table 2.3). A comparison of the model performance of 1960–80 with the performance 1980–00 would be required to answer the question whether the data quality is sufficient or not.

3.6.2 The Response of the Lake to Increased Air Temperature and Cloud Cover Fraction

High sensitivity of a model that is validated and verified for many different forcing conditions also indicates high sensitivity of the real lake system itself. This may not be true for models with a bad parameterisation of the physical processes. The less empirical the model and the higher the level of process description, the more valid is it to deduce a lake response from a model response.

If it is believed that a change in DYRESM’s simulation result due to a change in meteorological forcing is generally transferable to the real system, then the model has predictive capabilities. In light of an expected climate change, this is an important ability. The simulations with increased air temperature and increased cloud cover fraction are compared in figs. 3.20 and 3.21, respectively, to the unmodified simulations.

The higher air temperature increases the water temperature for the whole depth region 0-30 m and all seasons except for short periods of slightly decreased water temperatures. The temperature of the surface region increases more than below. This means stratification
becomes more stable. Lastly, stratification commences earlier than in the unmodified case, which is best seen in year 1982 where stratification begins with the end of April instead of mid of May. Also important is a trend to increasing water temperatures over the course of these three years. This could be due to the response time of the lake to the changed forcing that is longer than three years. To enable the lake system to achieve some new equilibrium the simulation periods must be much longer.

The case of increased cloud cover gives almost the inverse picture of the case of increased air temperature. Apart from a few spots below 20 m all water temperatures decrease. As the temperatures of the surface regions decrease more than below, the stratification becomes weaker. Wind events can more easily destroy stratification as seen at the beginning of July 1980. A slight trend towards continuously decreasing temperatures is also visible.

These contour plots can only provide a rough idea of what a change in one single forcing variable may mean for the lake system. In the scenarios of global warming many forcing variables change at the same time. This can lead to either counteracting single effects resulting in a negligible net effect or it can have effects that add up to an enormous net effect. Additionally, DYRESM would need to be coupled to CAEDYM as the forcing influences water temperatures, and through mixing, activity light and nutrient availability, which are all growth factors for algae. Algal distribution and concentration in turn influences the heat distribution and therefore the hydrodynamics.
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Figure 3.20: Temperature contour plots of a simulation with increased air temperature by 2 °C (top), a simulation with unmodified air temperatures (middle) and the water temperature differences (=modified minus unmodified) between the simulations (bottom). WMF=1.45, $C_{LN} = 1500$, PLT=0.5–2.5 m, $k_{ln} = 0.35$ m$^{-1}$, model start 15-Jan-1980. (Contouring grid: $dt=2$ days, $dz=2$ m, linear interpolation). For a colour version see fig. A.4.

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Figure 3.21: Temperature contour plots of a simulation with increased cloud cover fraction by 0.10 (top), a simulation with unmodified cloud cover fraction (middle) and the water temperature differences (=modified minus unmodified) between the simulations (bottom). WMF=1.45, $C_{LN} = 1500$, PLT=0.5–2.5 m, $k_{te} = 0.35$ m$^{-1}$, model start 15-Jan-1980. (Contouring grid: as fig. 3.20). For a colour version see fig. A.5.
Chapter 4

Conclusions

The standard code of DYRESM lacks a process based algorithm for diffusion in the hypolimnion and metalimnion. Simulations with that code proved unsatisfactory at Lake Constance. The thermocline was modelled with too large temperature gradients in the metalimnion resulting in an overestimation of temperatures above the thermocline and the opposite below it.

A DYRESM version with a diffusion algorithm that is still under development was then applied to Lake Constance. The conceptual idea of the algorithm that turbulent diffusion in the metalimnion is inversely proportional to the Lake Number is derived from experimental results at Lake Kinneret. The Lake Number can be considered in this context as a measure of internal wave activity. However, the level of process description is not yet high enough to avoid calibration of the constant of proportionality $C_{LN}$ in the Lake Number relationship, which might be a function of basin geometry. Future research may lead to the parameterisation of that “constant”.

A calibration of $C_{LN}$ in conjunction with the wind multiplication factor WMF and the maximum permissible layer thickness $PLT_{\text{max}}$ was performed on the two 3-year periods 1980–82 and 1995–97. The first period was deemed to represent a high trophic level of the lake and the second a low trophic level with accordingly different algal composition and concentration. The model performance was expressed objectively by three statistical values: the mean temperature error, the standard deviation and the root mean squared value of the temperature errors. The calibration had three important outcomes.

1. It confirmed a model sensitivity to $PLT_{\text{max}}$ suspected earlier. An explanation could not be given. With a thorough review of the algorithms and simple test cases that are analysed every model time step an answer might be found.

2. The calibration suggests the multiplication of wind speeds measured at the shore stations by a factor of 1.40–1.50. This agrees with estimates found in the literature. It would be interesting to know whether the wind data collected on eight lake stations by Appt (2002) during 5 weeks in autumn 2001 can confirm this kind of relationship between wind at the shore and wind across the lake.

3. A value of about 1500 is suggested for $C_{LN}$. This is approximately 5 times higher than what was found for Lake Kinneret and from model application on Mundaring Weir in Western Australia. This could be a consequence of the special geometry of Lake Constance with the long and narrow sub-basin Lake Überlingen which is separated by the sill of Mainau from the much larger main basin. Vertical diffusion coefficients near the sill were reported to be magnitudes higher than in the main basin so that the vertical flux at the sill may surmount the flux of the entire remaining lake in spite of the small area.

The verification of the parameters obtained by the calibration was done for an 11-year
CHAPTER 4. CONCLUSIONS

The model performance was very different from year to year and for most years worse than during the calibration periods. However, it was shown by an example that some of the extreme deviations may be attributed to the displacement of isotherms by internal waves. In a year with only about 12 field data profiles, one or two such cases will falsify the otherwise decent model performance. It shows that field data in much higher spatial and temporal resolution are required to gain confidence in a prospective lake management system with sophisticated (3D)-model applications. Nevertheless, after the calibration of the diffusion algorithm, DYRESM can be considered suitable for further research with the water-quality model CAEDYM.

The calibration was done with the assumptions that the yearly depth and time averaged light extinction coefficient \( k_{le} \) is 0.35 m\(^{-1}\) and that the model is not sensitive to the minimum permissible layer thickness \( PLT_{\text{min}} \) although the opposite was the case for \( PLT_{\text{max}} \). The value for \( k_{le} \) was determined from secchi-disk data as well as from monthly means of phytoplankton concentrations. Even though the large temporal and spatial variability of this parameter is a strong argument to calibrate \( k_{le} \), it was not considered for calibration because future model applications on Lake Constance should use the ecosystem model DYRESM-CAEDYM. In that case \( k_{le} \) is computed within the model as a function of depth and time making a calibration worthless. Instead, a sensitivity analysis was done for \( k_{le} \) and \( PLT_{\text{min}} \) to investigate their potential influence on the model performance if the optimal parameter set of \( PLT_{\text{max}} \), \( C_{LN} \) and WMF resulting from the calibration is used. The outcome was a weak model sensitivity to \( k_{le} \) and a negligible sensitivity to \( PLT_{\text{min}} \). Therefore the assumption not to consider these two parameters in the calibration was justified.

In light of an expected climate change, a sensitivity analysis was also done on two meteorological variables. Air temperature and cloud cover fraction were varied by 2 °C and 0.10 respectively. Both modifications had tremendous impact on the model performance. The analysis indicated better model performance for reduced short wave radiation, i.e. more cloud cover fraction. Excluding all other uncertainties this might suggest that the radiation measurements at elevated shore stations may not represent the radiation received at lake surface level where frequent fog or haze reduces radiation energy.

Under the assumption that DYRESM has predictive capabilities due to its high level of process description, model simulations can be compared under different meteorological forcing. This was done with a scenario of increased air temperature and for a scenario of increased cloud cover fraction for the 3-year period 1980–82. The case of increased air temperature by 2 °C led to an increase of average water temperature of almost 1 °C between 0–30 m depth over the unmodified case. Furthermore, stratification became stronger and longer in duration. However, a 3-year simulation was too short for the lake to attain an equilibrium with respect to the new environmental conditions. A longer run could not be done due to stability problems with the developmental code. The scenario of increased cloud cover by 0.1 resulted in the essentially opposite picture of the case of increased air temperatures. This is because higher cloud cover reduces incident short wave radiation.

Convincing statements about potential effects of global warming would, first of all, require to know the most likely change in meteorological forcing. This information could come from a regional atmospheric model that receives boundary conditions from global climate change model. This would result in a much more complex modification of the input than in the above examples. Secondly, the ecosystem model DYRESM-CAEDYM should be used to capture the feedback from phytoplankton on the hydrodynamics and vice versa. Lastly the use of CAEDYM might allow the simulation of ecosystem transitions due to global warming, e.g. change in phytoplankton composition.
Appendix A

Colour Figures
Figure A.1: Temperature contour plots of a simulation without the diffusion algorithm (top), IGKB field data (middle) and the temperature differences between simulation and field data (bottom). Dots indicate measurements. The wind multiplication factor (WMF) is 1.3, the permissible layer thickness (PLT) is 0.5–2.5 m, light extinction coefficient $k_{ie}$ is $0.35 \text{ m}^{-1}$ and the model was started on 15-Jan-1980. (Contouring grid: $dt = 2$ days for simulation, $dt = 4$ days for field data and differences, $dz = 2$ m, linear interpolation). Corresponds to fig. 3.1
DYRESM simulation: wmf1.45ln1500plt0.50_2.5le0.3519800115_19830111_0.35_TcGrd_t22.

IGKB Field Data. Fld_TCntrGrd_t4z2_1980001_1983001.

error = DYRESM minus field data. wmf1.45ln1500plt0.50_2.5le0.3519800115_19830111_error_TcGrd_t4z2.

Figure A.2: Temperature contour plots of a simulation with optimal parameters \( PLT_{max} = 2.5 \) m, \( C_{LN} = 1500 \) and \( WMF = 1.4 \) from calibration (top), IGKB field data (middle) and the temperature differences between simulation and field data (bottom). Dots indicate measurements. \( PLT_{min} = 0.5 \) m, \( k_{le} = 0.35 \) m\(^{-1}\) and the model was started on 15-Jan-1980. (Contouring grid: as fig. A.1). Corresponds to fig. 3.11
Figure A.3: Temperature contour plots of the period of worst model performance (top), IGKB field data (middle) and the temperature differences between simulation and field data (bottom). Dots indicate measurements. WMF=1.45, $C_{LN} = 1500$, PLT=0.5-2.5 m, $k_e = 0.35$ m$^{-1}$, model start 14-Feb-1984. (Contouring grid: as fig. A.1). Corresponds to fig. 3.18
Figure A.4: Temperature contour plots of a simulation with increased air temperature by 2 °C (top), a simulation with unmodified air temperatures (middle) and the water temperature differences (=modified minus unmodified) between the simulations (bottom). WMF=1.45, $C_{LN}=1500$, PLT=0.5–2.5 m, $k_{te}=0.35$ m$^{-1}$, model start 15-Jan-1980. (Contouring grid: dt=2 days, dz=2 m, linear interpolation). Corresponds to fig. 3.20.
Figure A.5: Temperature contour plots of a simulation with increased cloud cover fraction by 0.10 (top), a simulation with unmodified cloud cover fraction (middle) and the water temperature differences (=modified minus unmodified) between the simulations (bottom). WMF=1.45, $C_{LN} = 1500$, PLT=0.5–2.5 m, $k_\epsilon = 0.35$ m$^{-1}$, model start 15-Jan-1980. (Contouring grid: as fig. A.4). Corresponds to fig. 3.21.
Appendix B

Abbreviations, Symbols and Indices
### B.1 List of Abbreviations

<table>
<thead>
<tr>
<th>abbreviation</th>
<th>meaning</th>
</tr>
</thead>
<tbody>
<tr>
<td>a.s.l.</td>
<td>above sea level</td>
</tr>
<tr>
<td>BBL</td>
<td>benthic boundary layer</td>
</tr>
<tr>
<td>CCF</td>
<td>cloud cover fraction</td>
</tr>
<tr>
<td>CET</td>
<td>Central European Time</td>
</tr>
<tr>
<td>DWD</td>
<td>Deutscher Wetterdienst, Offenbach, Germany</td>
</tr>
<tr>
<td>DYRESM</td>
<td>Dynamic Reservoir Simulation Model</td>
</tr>
<tr>
<td>CAEDYM</td>
<td>Computational Aquatic Ecosystem Dynamics Modell</td>
</tr>
<tr>
<td>CWR</td>
<td>Centre for Water Research, The University of Western Australia, Perth</td>
</tr>
<tr>
<td>FOWG</td>
<td>Bundesamt für Wasser und Geologie BWG, Switzerland. (The Federal Office for Water and Geology FOWG)</td>
</tr>
<tr>
<td>HVZ</td>
<td>Hochwasser-Vorhersage-Zentrale (Department of LfU)</td>
</tr>
<tr>
<td>IES</td>
<td>UNESCO International Equation of State for seawater</td>
</tr>
<tr>
<td>IGKB</td>
<td>Internationale Gewässerschutzkommission für den Bodensee</td>
</tr>
<tr>
<td>ISF</td>
<td>Institut für Seenforschung (Department of LfU)</td>
</tr>
<tr>
<td>LfU</td>
<td>Landesanstalt für Umweltschutz LfU, Baden-Württemberg, Germany (The State Institute for Environmental Protection, Baden-Württemberg)</td>
</tr>
<tr>
<td>LWBA</td>
<td>Hydrographischer Dienst Vorarlberg, Landeswasserbauamt Bregenz, Austria</td>
</tr>
<tr>
<td>MCH</td>
<td>MeteoSchweiz (MeteoSwiss)</td>
</tr>
<tr>
<td>KE</td>
<td>Kinetic energy (of mean velocity)</td>
</tr>
<tr>
<td>PLT</td>
<td>permissible layer thickness</td>
</tr>
<tr>
<td>PSS</td>
<td>practical salinity scale</td>
</tr>
<tr>
<td>RMS</td>
<td>root mean squared, i. e. $\text{rms}(x_i) = \sqrt{1/n \sum_{i=1}^{n} x_i^2}$</td>
</tr>
<tr>
<td>SWR</td>
<td>short wave radiation</td>
</tr>
<tr>
<td>TKE</td>
<td>Turbulent kinetic energy</td>
</tr>
<tr>
<td>UILV</td>
<td>Umweltinstitut des Landes Vorarlbergs, Austria (note: UILV is not an official acronym)</td>
</tr>
<tr>
<td>UTC</td>
<td>Universal Time Coordinate (former Greenwich Mean Time GMT)</td>
</tr>
<tr>
<td>WMF</td>
<td>Wind multiplication factor</td>
</tr>
<tr>
<td>ZAMG</td>
<td>Zentralanstalt für Meteorology und Geodynamik, Austria (Central Institute of Meteorology and Geodynamics)</td>
</tr>
<tr>
<td>ZVBWV</td>
<td>Zweckverband Bodensee-Wasserversorgung, Germany</td>
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</table>
## B.2 Lists of Symbols

### B.2.1 Roman Symbols

<table>
<thead>
<tr>
<th>symbol</th>
<th>common units</th>
<th>meaning</th>
</tr>
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<tbody>
<tr>
<td>$a$</td>
<td>any</td>
<td>fitting constants from regression analysis</td>
</tr>
<tr>
<td>$A$</td>
<td>any</td>
<td>a amplitude</td>
</tr>
<tr>
<td>$A_m$</td>
<td>$m^2$</td>
<td>area</td>
</tr>
<tr>
<td>$AST$</td>
<td>(hr)</td>
<td>apparent solar time</td>
</tr>
<tr>
<td>$B_N$</td>
<td></td>
<td>the Burger number</td>
</tr>
<tr>
<td>$c$</td>
<td>kg/m$^3$</td>
<td>a concentration</td>
</tr>
<tr>
<td>$C$</td>
<td></td>
<td>bulk aerodynamic transfer coefficients and other constants</td>
</tr>
<tr>
<td>$C_{eva}$</td>
<td>kJ/kg</td>
<td>Latent heat of evaporation of water $C_{eva} = 2500$ kJ/kg</td>
</tr>
<tr>
<td>$C_P$</td>
<td>kJ/(kg-K)</td>
<td>heat capacity of water at constant pressure $C_P = 4.19$ kJ/(kg-K)</td>
</tr>
<tr>
<td>$d$</td>
<td>m</td>
<td>depth</td>
</tr>
<tr>
<td>$CCF$</td>
<td></td>
<td>cloud cover fraction</td>
</tr>
<tr>
<td>$CET$</td>
<td>(hr)</td>
<td>Central European Time</td>
</tr>
<tr>
<td>$F$</td>
<td></td>
<td>Froude Number</td>
</tr>
<tr>
<td>$g'$</td>
<td>m/s$^2$</td>
<td>$g' = \Delta \rho/\rho \cdot g$, with $g$ as acceleration of gravity</td>
</tr>
<tr>
<td>$G$</td>
<td>N</td>
<td>gravity force</td>
</tr>
<tr>
<td>$h$</td>
<td>m</td>
<td>thickness</td>
</tr>
<tr>
<td>$h_r$</td>
<td>%</td>
<td>relative humidity</td>
</tr>
<tr>
<td>$H$</td>
<td>m</td>
<td>water level</td>
</tr>
<tr>
<td>$e$</td>
<td>mbar</td>
<td>vapour pressure</td>
</tr>
<tr>
<td>$E$</td>
<td>J</td>
<td>a energy</td>
</tr>
<tr>
<td>$JD$</td>
<td>day</td>
<td>the day of a year: 1. Jan. is $JD = 1$; 31. Dec. is $JD = 365,366$</td>
</tr>
<tr>
<td>$k$</td>
<td>any</td>
<td>constants</td>
</tr>
<tr>
<td>$L$</td>
<td>m</td>
<td>a length</td>
</tr>
<tr>
<td>$L_N$</td>
<td></td>
<td>the Lake Number</td>
</tr>
<tr>
<td>$KE$</td>
<td>J</td>
<td>kinetic energy (of mean velocity)</td>
</tr>
<tr>
<td>$n$</td>
<td></td>
<td>a number</td>
</tr>
<tr>
<td>$q$</td>
<td>W/m$^2$</td>
<td>a energy flux</td>
</tr>
<tr>
<td>$Q$</td>
<td>m$^3$/s</td>
<td>a discharge</td>
</tr>
<tr>
<td>$Ri$</td>
<td></td>
<td>Gradient Richardson Number</td>
</tr>
<tr>
<td>$P$</td>
<td>mbar</td>
<td>ambient atmospheric pressure (966 mbar at Lake Constance)</td>
</tr>
<tr>
<td>$r$</td>
<td></td>
<td>a reflection coefficient (albedo)</td>
</tr>
<tr>
<td>$S$</td>
<td>(PSS78)</td>
<td>salinity</td>
</tr>
<tr>
<td>$S$</td>
<td>N/m</td>
<td>stratification strength</td>
</tr>
<tr>
<td>$t$</td>
<td>s, day</td>
<td>a time</td>
</tr>
<tr>
<td>$T$</td>
<td>s, day</td>
<td>a time period</td>
</tr>
<tr>
<td>$TKE$</td>
<td>J</td>
<td>turbulent kinetic energy</td>
</tr>
<tr>
<td>$u$</td>
<td>m/s</td>
<td>water velocity</td>
</tr>
<tr>
<td>$U$</td>
<td>m/s</td>
<td>wind velocity</td>
</tr>
<tr>
<td>$u_*$</td>
<td>m/s</td>
<td>friction velocity $u_* = \sqrt{\tau/\rho}$</td>
</tr>
<tr>
<td>$v_i$</td>
<td>m/s</td>
<td>phase speed of a internal wave</td>
</tr>
<tr>
<td>$V$</td>
<td>m$^3$</td>
<td>volume</td>
</tr>
<tr>
<td>$z$</td>
<td>m</td>
<td>vertical coordinate</td>
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## B.2.2 Greek Symbols

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<thead>
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<th>symbol</th>
<th>common units</th>
<th>meaning</th>
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<tr>
<td>$\alpha$</td>
<td>$^\circ$</td>
<td>half the base angle of the triangular shaped river cross-section</td>
</tr>
<tr>
<td>$\beta$</td>
<td>rad</td>
<td>a phase shift or an angle</td>
</tr>
<tr>
<td>$\gamma$</td>
<td>–</td>
<td>photofraction</td>
</tr>
<tr>
<td>$\delta$</td>
<td>rad</td>
<td>declination of the sun at the equator</td>
</tr>
<tr>
<td>$\epsilon$</td>
<td>–</td>
<td>an emissivity</td>
</tr>
<tr>
<td>$\epsilon$</td>
<td>any</td>
<td>an error term</td>
</tr>
<tr>
<td>$\varepsilon$</td>
<td>–</td>
<td>an entrainment coefficient constant</td>
</tr>
<tr>
<td>$\zeta$</td>
<td>any</td>
<td>a deviation</td>
</tr>
<tr>
<td>$\eta$</td>
<td>–</td>
<td>a efficiency</td>
</tr>
<tr>
<td>$\theta$</td>
<td>K, °C</td>
<td>(absolute) temperature</td>
</tr>
<tr>
<td>$\kappa$</td>
<td>$\mu$S/cm</td>
<td>electric conductivity</td>
</tr>
<tr>
<td>$\phi$</td>
<td>$^\circ$</td>
<td>longitudinal slope of river bed downstream of mouth</td>
</tr>
<tr>
<td>$\phi$</td>
<td>$^\circ$</td>
<td>latitude, positive to the north</td>
</tr>
<tr>
<td>$\lambda$</td>
<td>$^\circ$</td>
<td>longitude, positive to the east</td>
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<tr>
<td>$\rho$</td>
<td>kg/m$^3$</td>
<td>mass density</td>
</tr>
<tr>
<td>$\sigma$</td>
<td>W/(m$^2$·K$^4$)</td>
<td>Stefan-Boltzmann constant $\sigma = 5.6696 \cdot 10^{-8}$ W/(m$^2$·K$^4$)</td>
</tr>
<tr>
<td>$\tau$</td>
<td>N/m$^2$</td>
<td>(wind) stress</td>
</tr>
<tr>
<td>$\chi$</td>
<td>–</td>
<td>normalised sensitivity coefficient</td>
</tr>
<tr>
<td>$\psi$</td>
<td>–</td>
<td>a fraction</td>
</tr>
<tr>
<td>$\omega$</td>
<td>1/s</td>
<td>a frequency</td>
</tr>
<tr>
<td>$\Omega$</td>
<td>deg/hr</td>
<td>angular velocity of the earth rotation</td>
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### B.3 List of Indices

<table>
<thead>
<tr>
<th>index</th>
<th>meaning</th>
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<tbody>
<tr>
<td>a</td>
<td>air; albedo</td>
</tr>
<tr>
<td>BBL</td>
<td>benthic boundary layer</td>
</tr>
<tr>
<td>bg</td>
<td>background</td>
</tr>
<tr>
<td>bp</td>
<td>buoyant plume</td>
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<tr>
<td>corr</td>
<td>correction</td>
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<td>drag</td>
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<td>effective</td>
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<td>eva</td>
<td>evaporation</td>
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<td>G</td>
<td>gravity</td>
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<tr>
<td>IES</td>
<td>(UNESCO) International Equation of State</td>
</tr>
<tr>
<td>inf</td>
<td>inflow(s)</td>
</tr>
<tr>
<td>int</td>
<td>interior</td>
</tr>
<tr>
<td>KE</td>
<td>kinetic energy (of mean velocity)</td>
</tr>
<tr>
<td>L</td>
<td>latent heat</td>
</tr>
<tr>
<td>LC</td>
<td>Lake Constance</td>
</tr>
<tr>
<td>le</td>
<td>light extinction</td>
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<td>lw</td>
<td>long wave</td>
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<tr>
<td>M</td>
<td>momentum</td>
</tr>
<tr>
<td>mix</td>
<td>mixed (volume)</td>
</tr>
<tr>
<td>mol</td>
<td>molecular</td>
</tr>
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<td>P</td>
<td>pressure</td>
</tr>
<tr>
<td>PE</td>
<td>potential energy</td>
</tr>
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<td>phytoplankton</td>
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<td>precipitation</td>
</tr>
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<td>rg</td>
<td>re-gridding</td>
</tr>
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<td>s</td>
<td>saturation</td>
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<td>S</td>
<td>sensible heat</td>
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<tr>
<td>sd</td>
<td>secchi-disk</td>
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<td>shear flow</td>
</tr>
<tr>
<td>sh</td>
<td>(time) shift</td>
</tr>
<tr>
<td>sp</td>
<td>shear period</td>
</tr>
<tr>
<td>sw</td>
<td>short wave</td>
</tr>
<tr>
<td>tc</td>
<td>thermocline</td>
</tr>
<tr>
<td>w</td>
<td>water</td>
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<tr>
<td>ws</td>
<td>wind stirring</td>
</tr>
<tr>
<td>wdr</td>
<td>withdrawal</td>
</tr>
<tr>
<td>z</td>
<td>vertical</td>
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ANTENUCCI, J. 2001 The coupled CWR Dynamic Reservoir Simulation Model and Computational Aquatic Ecosystem Dynamics Model DYRESM-CAEDYM. User-Guide. Centre for Water Research, The University of Western Australia.


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- Umweltinstitut des Landes Vorarlbergs (Dietmar Buhmann)
- Federal Office for Water and Geology, Landeshydrologie (Daniel Streit)
- Landesanstalt für Umweltschutz Baden-Württemberg (Bernd Wahl from ISF and Werner Schulz from HVZ)
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