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Vertical structure of currents in western Lake Constance

Bertram Boehrer [1]

UFZ - Centre for Environmental Research Ltd., Department of Inland Water Research Magdeburg, Germany

Johann Ilmberger and Karl Otto Münnich

Institut für Umweltphysik, Universität Heidelberg, Heidelberg, Germany

Abstract. Simultaneous measurements of current velocity and density profiles at the Sill of Mainau (western Lake Constance) enabled the observation of the internal response of the lake and the calculation of gradient Richardson number over the entire water column at the sampling site. Periodic changes of current direction corresponded with predicted periods of the two-layer oscillation and could be confirmed with thermocline oscillation data. In general, the current profiles confirmed a two-layer structure. However, an activating wind of speeds higher than 4 m/s produced three-layer profiles for a period of several hours at the sampling site, when the velocities due to the two-layer oscillation traversed zero. At times the implied current shear produced supercritical values of gradient Richardson number on a vertical scale of a few meters. This occurred during winds in the epilimnion and after strong (> 4 m/s) winds in the lower thermocline and the hypolimnion. In the 20 m above the lake bed, topographical conditions helped the gradient Richardson number to go supercritical sporadically. Across the thermocline, no supercritical gradient Richardson numbers could be confirmed at the measuring site. A lower boundary for vertical mixing in the measuring site due to current shear could be evaluated.

1. Introduction

Lake Constance (in German, Bodensee) in central Europe can be classified as a deep lake, and with its volume of 48.5 km^3 it may be called a small, large lake. It is located at the northern edge of the Alps at a latitude of 47°N. We consider the upper two parts (Figure 1): Obersee contains 94% of the water and is 253 m deep. Connected with it across the 99 m deep Sill of Mainau is the relatively narrow and deep (145 m) Überlinger See. (For details on the hydrological aspects, see *Bäuerle et al.* [1998]) In some long and deep lakes, current velocities have previously been measured, but either only a small portion of a seiche period [*Thorpe*, 1977] was covered, or only 3 to 5 current meters were used over the entire depth [e.g. *Hollan and Simons*, 1978; *Hamblin and Hollan*, 1978; *Hutter*, 1986].

According to the simple two-layer model, a stratified lake reacts to a wind with an internal oscillation [*Wedderburn*, 1907; *Mortimer*, 1952; *Thorpe*, 1972; *Zenger*, 1989]. Two layers separated by the thermocline flow in opposite directions and reverse periodically. The period of the oscillation can be estimated and comparison with field data generally yields a good agreement (for Lake Constance, see *Zenger* [1989]). Numerical models have successfully reproduced the two-layer oscillation also including the oscillations of horizontal higher order [*Bäuerle*, 1981; *Zenger*, 1989; *Bauer et al.*, 1994; *Hertkorn*, 1996; *Hutter et al.*, 1998].

Also more-layer oscillations can be considered and their oscillation periods can be evaluated; however, nature shows a dominant first baroclinic mode (two-layer oscillation) in most lakes. Frequency analyses of temperature time series yield smaller contributions of higher modes. In only few lakes a higher mode prevails [*Wiegand and Chamberlain*, 1987; *Münnich et al.*, 1992].

Yet only little is known about the thickness of the transition zone between the flowing layers. Additionally, it is a challenge to investigate the current profile at the moment of the direction change of the predominant first baroclinic mode. If the theoretical considerations of interfering waves of various baroclinic modes is correct, the higher modes should show up, if their

[1] Previously at Institut für Umweltphysik, Heidelberg, Universität Heidelberg, Germany.

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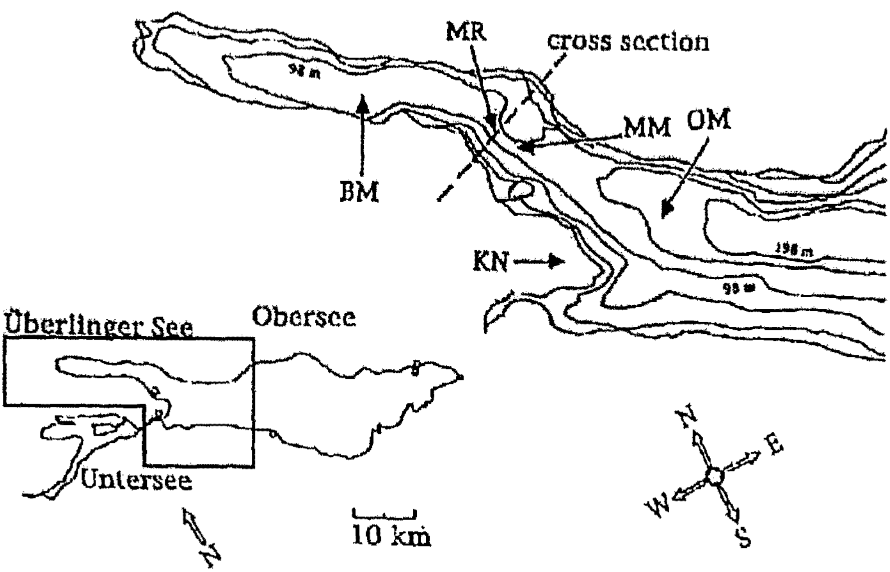

Figure 1. The Western part of Lake Constance including the location of the sampling stations (double capitals) and the cross section (Figure 2) which forms the dividing line between the lake parts Überlinger See and Obersee.

contribution is large enough. Lake Constance's large depth, as well as its elongated shape and its steep sidewalls make it a well-suited laboratory for clearly defined velocity profiles.

Sills and narrows can enhance current velocities and restrict flows to a smaller vertical scale and thus imply higher velocity shear in a water column. Hence it has been assumed and shown that they can be responsible for much of the vertical transport in a water body [e.g., *Kocsis et al.*, 1998]. A prominent candidate for investigating these effects in Lake Constance is the Sill of Mainau which separates the wider basin of Obersee from the narrower channel of Überlinger See. Vertical transport coefficients in the lower thermocline [*Heinz*, 1990; *Heinz et al.*, 1990] are enhanced in Lake Constance and measurements of the turbulent structure of the water column at the Sill of Mainau could prove that dissipation of the internal wave energy is enhanced within this region, contributing about 40% of the basin-wide dissipation [*Kocsis et al.*, 1998].

In this contribution we present current profile measurements in Lake Constance at the Sill of Mainau far off the shoreline with a high resolution over the entire depth. Four events of an excited internal response of Lake Constance have been covered with current velocity profiles. From simultaneous current velocity profiles and temperature profiles we evaluate gradient Richardson numbers and confirm supercritical values on a scale of several meters and describe the spatial and temporal pattern. We try to clarify, whether at all and at what depths the shear induced by the internal waves can contribute considerably to the vertical transport through the stratified water column.

2. Measurements

All sampling sites were named with double capitals (KN, BM, MR, and MM in Figure 1). At BM, thermistor chains were deployed. Sampling site MM carried two current meters in depths of 25 and 80 m. At KN the "Deutscher Wetterdienst" ran a meteorological station, which delivered the information about the wind speed in this article. For a closer description of any of these sampling sites and the deployed probes, we refer to *Heinz* [1990].

The profile current meter (PCM, Aanderaa, Norway) was deployed at MR. The site was on the shallowest part of the connection of Obersee and Überlinger See, where the water depth measured about 99 m. An anchor was set at 1100 m from the shoreline (Figure 2) with a pulley mounted which led a rope. The lake side end of the rope was tied to the PCM which was carried by a buoy (together they applied a force of about 200 N in the rope), while its other end was tied to a winch on the shoreline. This mechanism allowed sampling depths between 5 m below the surface and 1 m above the bed.

The PCM carried sensors for pressure, temperature, current direction, and current speed. For the current direction a magnetic compass was used, and the speed was measured mechanically. All sensors were read at a sampling rate of 30 s and the data were transferred to an electronic memory.

Usually, the PCM was pulled deeper at intervals of 1 min, synchronously with the sampling periods. The measurements during the pulling action were removed, leaving only the data acquired while resting at constant depth. Sixty eight profiles were taken (64 of those

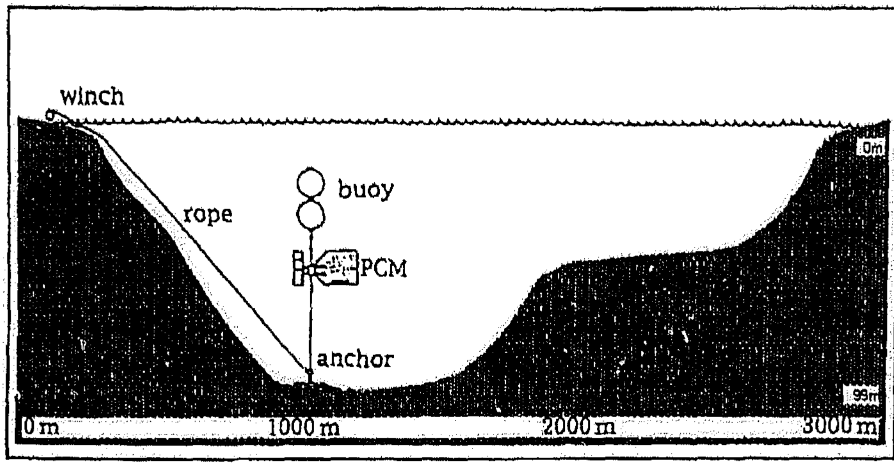

Figure 2. Cross section of Überlinger See through sampling site MR, including the construction of the profile current meter (PCM): current meter, anchor carrying a pulley, winch, and buoys.

from surface to bottom) in a field experiment between September 29, 1989 and November 9, 1989, all during autumnal temperature stratification (Figure 3) at intervals of 8 or 16 hours. The depth resolution was about 1.40 m, which resulted in 70 data points per profile and a duration of 70 min for the acquisition of most profiles.

A discussion of the accuracy of the velocity measurements was done by *Böhrer* [1990]. He considered the following sources of error: the threshhold of 2 cm/s, the accuracy of the current meter in standard use, and the impact of the vertical transport, for example transport against the current in the angle of the rope by the pulling action, possible pendulum movements, and hidden effects. In conclusion, he claimed an accuracy of 1.4 cm/s plus 30% of the velocity difference in the previous step. The temperature was measured by a sensor on top of the current meter. The absolute accuracy of the temperature measurements was of the order of 0.17 K. The resolution of the thermistor was 0.023 s with a response time of 70 s [*Böhrer*, 1990].

In Lake Constance the stratification was due to the temperature gradient. The electrical conductivity gradient indicated a much subordinate contribution due to dissolved ions, gaining some importance only below 75 m of depth. We included additional information from a conductivity-temperature-depth probe at a nearby station and calculated density using the formula by *Bührer and Ambühl* [1975] with a correction by *Heinz* [1990] (included in the work by *Bäuerle et al.* [1998]) for Lake Constance water. One example of the density profile was included in Figure 3. During the entire period of velocity measurements, thermistor data were acquired at BM. Their reading supplied the information for the depth of the 9°C isotherm shown in Figures 4 - 7, third panel.

3. Results and Discussion

After a wind a stratified lake reacts with an internal oscillation. Two layers separated by the thermocline

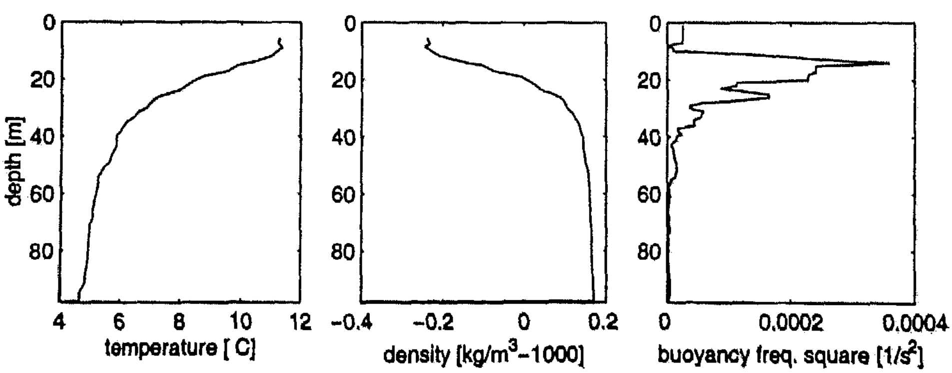

Figure 3. Autumn profiles (profile 40, October 12, 1989) of Lake Constance; density and Brunt-Väisälä-frequency calculated from the same measurements.

flow in opposite directions and reverse periodically. For many purposes, the Merian formula [e.g. *Mortimer*, 1952] is sufficient to estimate the oscillation period:

$$t = \frac{2L}{\sqrt{g \frac{\Delta\rho}{\rho} \frac{h_E h_H}{h_E + h_H}}}. \quad (1)$$

Including the acceleration due to Earth's gravity $g = 9.8\,\mathrm{m/s^2}$, and the values from Lake Constance of basin length $L = 60\,\mathrm{km}$, relative density difference $\Delta\rho/\rho = 0.8 \times 10^{-3}$, thickness of epilimnion and hypolimnion $h_E = 20\,\mathrm{m}$ and $h_H = 80\,\mathrm{m}$, we yielded an oscillation period of about 4 days, which confirmed results of previous studies [*Zenger*, 1989; *Heinz*, 1990; *Schimmele*, 1993].

3.1. Weak Wind Events and Two-Layer Oscillation

Two observations of weak wind events (Figure 4 and Figure 5) were made, both confirming the simple imagination of the two-layer model. The oscillation started after the exciting wind had released (Figure 4, first panel), and the thermocline showed a close to sinusoidal oscillation (third panel). Coinciding with the maxima and minima of the elevation of the thermocline, the current direction at 80 m depth in station MM changed abruptly (fourth panel). In the top panels we find the temporal locations of the velocity profiles displayed in the bottom panel as a staggered plot.

The reversal of current direction coincided with the thermocline elevation in BM. The period for a cycle was close to the predicted $t = 4$ days. The zero transition of the reversal of the current direction at the end of a half cycle happened smoothly, at least no unexpected features were observed. Neither did the wind blowing from a northerly to northeasterly direction from day 2.2 to 2.5 affect this simple structure. Considering an Ekman deflection to the right, the wind might have supported the epilimnion with its flow to the west. The results of a simple two-layer structure were confirmed by a second weak wind event (Figure 5).

3.2. Strong Wind Events and Three-Layer Profiles

After strong winds (Figure 6 and Figure 7, first panel), the oscillation could be confirmed with an oscillation period of about 4 days (third panel). The maxima and minima of the thermocline elevation coincided with the current direction reversal (fourth panel). Most of the profiles showed a two-layer structure (bottom panel).

The profile measured at the time of reversal of the current direction showed a clear three-layer structure (profile 37, Figure 6, bottom panel). This feature was clearly reproduced in the second strong wind event (Figure 7, bottom panel) by profiles 80, 81 and 82. The latter sequence could even prove that the feature persisted for as long as 6 hours.

The origin of the waters of intermediate density which flow out of Überlinger See in the three-layer profiles is not known reliably. The most probable process for the production is an upwelling at the upwind end during the wind event, as it was observed at other times [*Bäuerle et al.*, 1998]. Also propagation speed of a second-mode wave ($c_2 = 0.1\,\mathrm{m/s}$ [*Boehrer*, this issue]) and the distance upwind end to MR (15 km) coincided with 2 days pretty well. However, a quantitative evaluation of the contributing effects was not feasible with the present data. As indicated below, the current shear in MR did not suffice to make the gradient Richardson number go supercritical across the thermocline to enhance diapycnal mixing in the thermocline sufficiently.

3.3. Indication of Lateral Structure

The internal Rossby radius $r_R = c/f = 3.5\,\mathrm{km}$ for a two-layer oscillation using a phase velocity $c = 2L/t$, with $t = 4$ days the period of the two-layer oscillation, and the Coriolis parameter $f = 1.06 \times 10^{-4}/\mathrm{s}$ at latitude 47°N, was larger than the width $b = 2.5\,\mathrm{km}$, of Lake Constance at the Sill of Mainau. For higher mode oscillations [see *Boehrer*, this issue], the result for the Rossby radius would be considerably (and increasingly) smaller. This indicated a two-layer flow would be only slightly affected by Earth rotation in Überlinger See, while in Obersee effects should be visible. Because of decreasing phase speeds [*Boehrer*, this issue], second-mode waves in Überlinger See were expected to show rotational effects, while it was doubtful whether higher (than second) modes could exist in pelagic waters (MR) at all.

A close investigation of the lateral structure was beyond the scope of this paper; however, the data were suited to give some clues whether lateral structure was present. The net transport into the basin of Überlinger See included the fluxes through all boundaries. Neglecting groundwater flows, only the dividing cross section through MR had to be considered:

$$V_N = \oint \vec{v} \cdot d\vec{a} = \int_S \vec{v} \cdot d\vec{a} = -\int_{z=0\mathrm{m}}^{98\mathrm{m}} u(z)\, b(z)\, dz, \quad (2)$$

where $\vec{v}$ and u represented current velocity and current speed, $b(z)$ included the width of the cross section in MR as a function of depth. The results are listed in Table 1.

If the net transport yielded a higher value than surface seiche (16 cm from minimum to maximum water level in 27 min; *Hamblin and Hollan*, [1978]) and measurement accuracy would allow (together $1800\,\mathrm{m^3/s}$), some lateral structure had to be present. This was the case for early profiles (during the internal response after strong winds) as well as for the three-layer profiles. While for the former case a definite answer could not be evaluated, the latter might be attributed to rotational effects [see also *Boehrer*, this issue].

4. Gradient Richardson Number

The gradient Richardson number represents the ratio of inertial and buoyancy terms. *Miles* [1962] and

Howard [1962] proved that a shear flow can go unstable, if the gradient Richardson number falls below 0.25. In this section we derive the stability criterion from energy arguments and consider an overturning process a fully inelastic impact of two water parcels.

In a shear flow, two water parcels of finite thickness and equal size are considered. Relative to the center of mass, each parcel moves at a speed of $\Delta u/2$ (see Figure 8). Together their motion includes a kinetic energy of

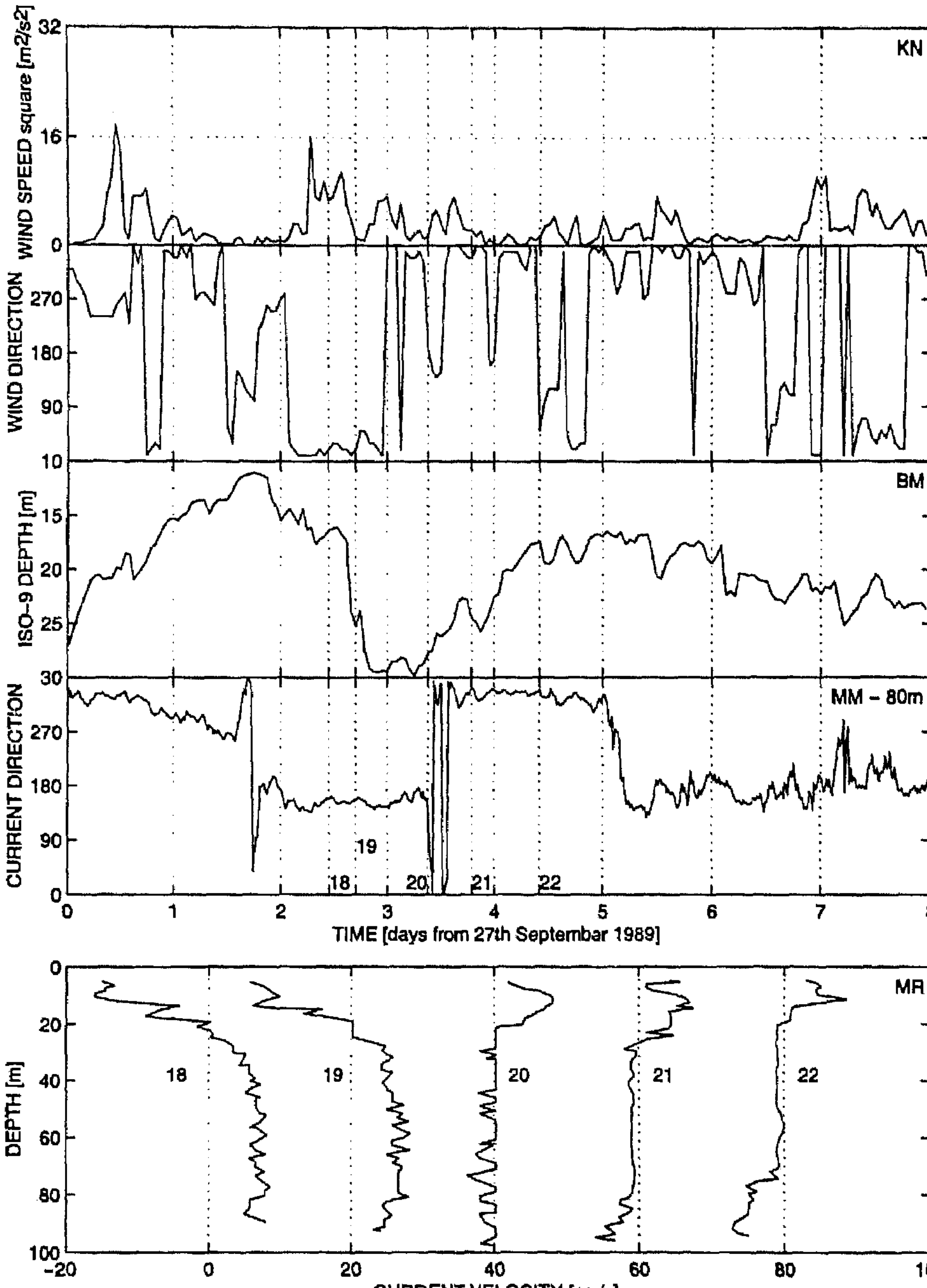

Figure 4. Internal response after September 27th, 1989 displayed by the wind speed, wind direction, the depth of the 9°C-isotherm and the current direction. Acquisition sites are marked in the upper right corner of the respective panel. The temporal location of each velocity profile was marked by a dotted line in the upper panels, while the bottom panel includes the velocity profiles in a staggered plot with an offset of 20 cm/s between profiles. Negative speeds indicate flow into Überlinger See.

$$E_k = 2 \cdot \frac{1}{2}\rho V\left(\frac{\Delta u}{2}\right)^2, \tag{3}$$

where ρ is the average density and V is the volume of the parcels (which is valid for a small relative density difference $\Delta\rho/\rho$).

By some process the parcels may be forced to exchange their position in the vertical. The only available energy source is the kinetic energy of the respective overturning water parcels. Potential energy would be gained:

$$E_p = gV\Delta\rho\Delta z, \tag{4}$$

where g is the acceleration due to gravity, $\Delta\rho$ is the density difference between the parcels, and finally Δz is the vertical distance of the centers of the water parcels.

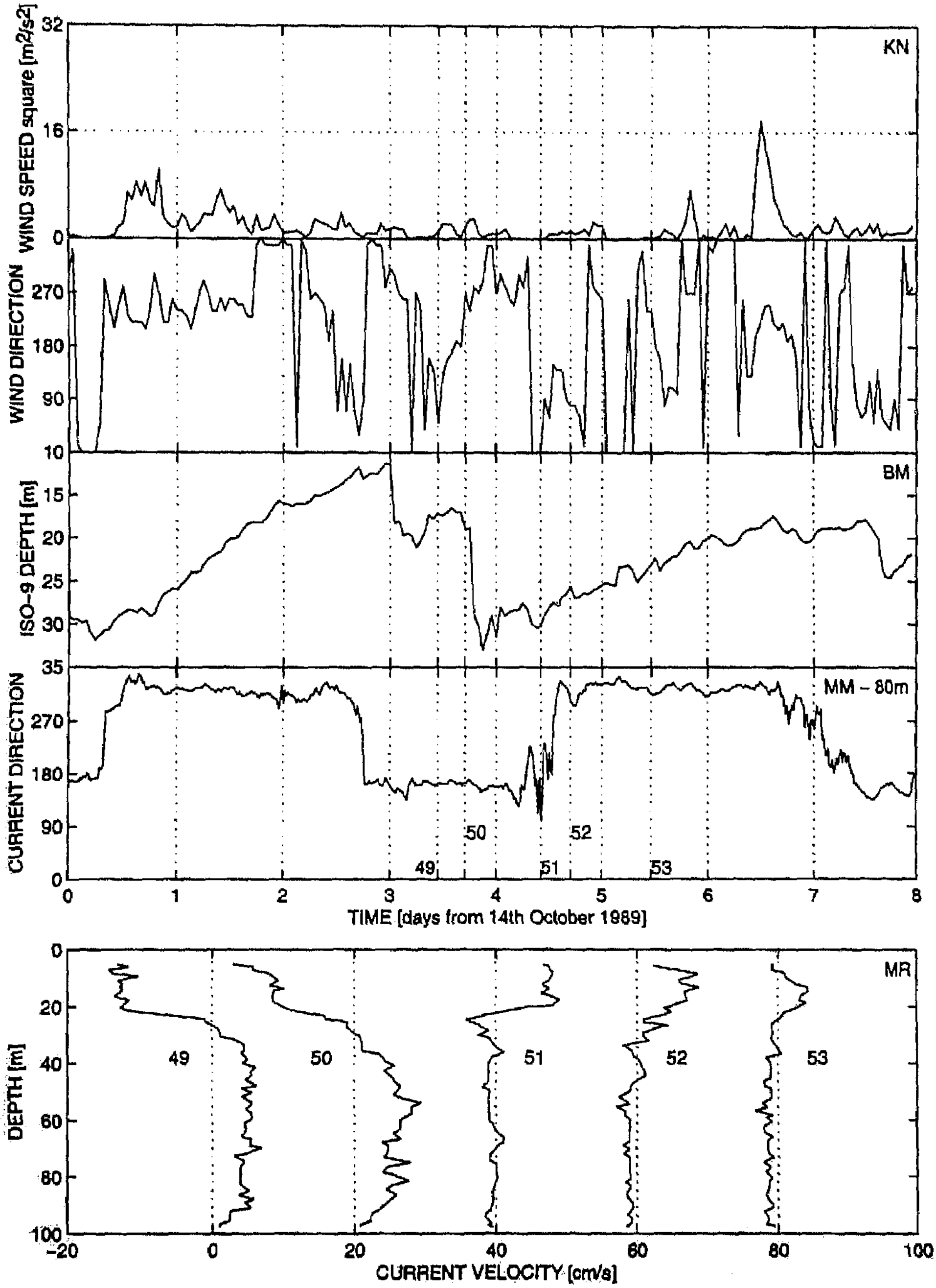

Figure 5. Internal response after October 14, 1989. For more details, see Figure 4 caption.

Table 1. Calculated Net Transport at Cross Section Through Measuring Site MR Based on the Assumption of no Lateral Structure

Profile	Net Transport m^3/s	Remarks
18	-2362	strong wind, early profile
35	-9900	strong wind, early profile
49	1181	weak wind, early profile
75	-1738	weak wind, early profile
19	-25	strong wind, late profile
36	2787	strong wind, late profile
50	756	weak wind, late profile
78	768	weak wind, late profile
37	-2526	strong wind, three-layer profile
80	-2647	strong wind, three-layer profile
52	-2278	weak wind, direction change

be calculated with any accuracy. Outside this contour the velocity gradients were too small and thus probably of minor importance.

In the hypolimnion, good results of Richardson numbers could be achieved during the internal response to a strong wind, where the results become denser (profiles 36 to 39 and profiles 82 and 83). Reliable evaluation of Richardson numbers was possible more frequently in the bottom 20 m where bottom stress might be felt, or the topography might imply higher gradients. The occasions when the upper bound of the Richardson number fell below critical (< 0.25) are marked dark in Figure 10. Only isolated events force supercritical Ri on a vertical length scale of several meters.

4.2. Temporal Pattern

Above the thermocline, supercritical Richardson numbers are found in profiles 32 to 34, and 68 to 74 during strong winds. Two days after the wind released, high velocity gradients were created by the three-layer profiles in the lower thermocline (30 to 40 m depth). After both events of a strong wind, Richardson numbers dropped below 0.25, probably only for a relatively short period of time, while the first-mode internal wave transversed zero. Later during the response, supercritical Richardson numbers were found in the hypolimnion.

Close to the lake bed (below 75 m depth), supercritical Richardson numbers appeared sporadically, without a clear temporal pattern, though events seemed to occur more densely during winds and toward the end of the modal response.

In all cases the internal response was excited by westerly winds, which were known to prevail at least during the stratification period. Unfortunately, the *Kocsis et al.* [1998] measurements took place during the rare occasion of NE winds, which were not covered by this approach. As a consequence, the Kocsis et al. measurements could not be directly inserted into the observed pattern.

4.3. Implication for Vertical Transport

The energy consideration above showed that for $Ri <$ 0.25, instabilities can be sustained from the energy implied by the current shear. As a consequence, below this critical value an additional vertical transport process is permitted, which naturally will result in a rapid change of vertical transport coefficients in the neighborhood of $Ri = 0.25$.

Modified by interaction with any kind of disturbance, we will see a more smooth transition in a natural waters. *Peters et al.* [1988] and *Münnich et al.* [1992] performed measurements to connect gradient Richardson number with vertical transport processes, in the equatorial undercurrent or an alpine lake, respectively. *Imboden and Wüest* [1995] combined the results in a plot from which we read the turbulent diffusivity versus gradient Richardson number as

$$K_d = 3\times10^{-9}Ri^{-9.6} + 7\times10^{-6}Ri^{-1.3} + 1.4\times10^{-7}\,[\mathrm{m^2/s}] \qquad \text{for } Ri > 0.22, \quad (6)$$

where we also added the molecular diffusion of heat.

This curve yielded roughly a value of $2\times10^{-3}\mathrm{m^2/s}$ at the critical value of $Ri = 0.25$. It was very improbable

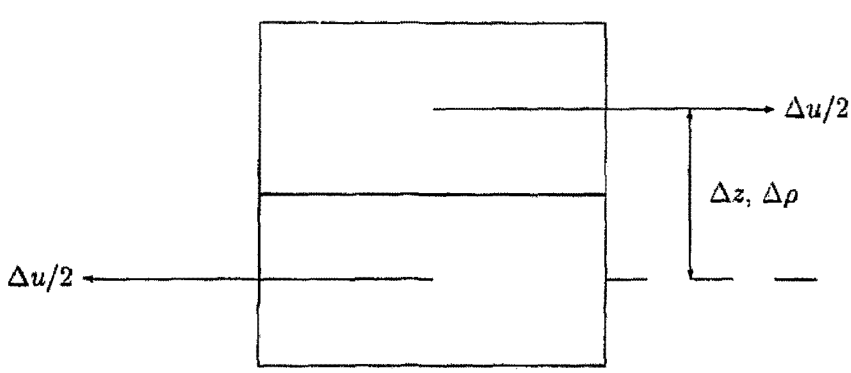

Figure 8. Two water parcels in a current shear.

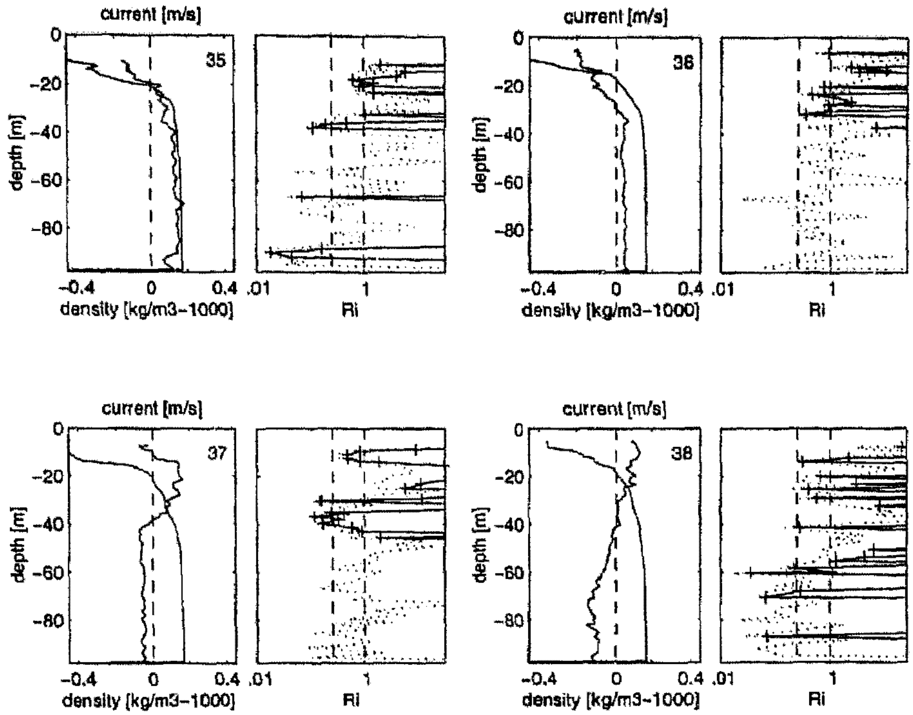

Figure 9. Profiles of gradient Richardson number in MR (dotted line), derived from simultaneous velocity (jagged line) and density profiles (smooth line) of the corresponding number. The plusses mark an upper bound for the Richardson number considering the accuracy of the measurement. Lines are drawn to guide the eye.

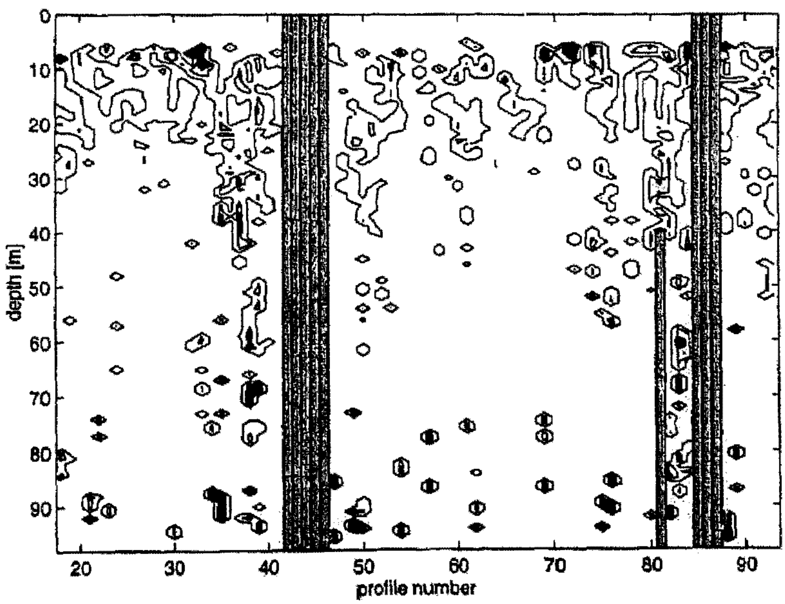

Figure 10. Contour plot of upper bound of the gradient Richardson number in measuring site MR. The outer contour comprise the area where an upper bound could be supplied. The shaded area indicates $0.25 < Ri < 1$, and the solid area indicates $Ri < 0.25$. Profiles with no current velocity measurements have been shaded on the plot.

that K_d kept on rising at the slope at $Ri^{-9.6}$ toward smaller Ri. In fact, the hourly averages displayed by *Peters et al.* [1988, Fig. 15] indicated a coefficient $2 \times 10^{-3}\,\mathrm{m^2/s} < K_d < 2 \times 10^{-2}\,\mathrm{m^2/s}$ for $0.1 < Ri < 0.25$. In conclusion, $K_d = 2 \times 10^{-3}\,\mathrm{m^2/s}$ for $0.1 < Ri < 0.25$ could be considered a lower boundary.

It could be shown that $K_d \leq 5 \times 10^{-2}\,\mathrm{m^2/s}$ for $0.1 < Ri < 0.25$: the assumption of $K_d = 5 \times 10^{-2}\,\mathrm{m^2/s}$ and claiming the velocity profiles at MR represented the conditions of at least 1 % of the area of Lake Constance ($\approx 0.5\,\mathrm{km^2}$), the vertical transport in this area surmounted the vertical transport within the entire lake, as from *Heinz* [1990] (see below).

The observations included 55 velocity profiles from the lake surface to the lake bed. We assumed they were distributed statistically over the entire period (at least within the required accuracy) and they were assumed to represent current shear over the entire period appropriately. Hence we took the average of the diffusion coefficient (calculated following (6)) over all profiles at all depths.

As the Richardson number, and thus the value for K_d, was evaluated from differences over a depth increment of the order of 5 m, we additionally displayed averages over 5 m in the vertical (Figure 11, solid line).

The vertical diffusivities were compared to the results by *Heinz* [1990] and *Heinz et al.* [1990] calculated with the gradient flux method each over a period of most of a summer: 1987, 1988, or 1989, respectively. Both approaches yielded a similar pattern. The vertical transport coefficients in Lake Constance showed a minimum at the depth of the thermocline (~20-30 m) with a value in the range of 10^{-6} to $10^{-5}\,\mathrm{m^2/s}$. In the epilimnion and in the hypolimnion the transport coefficients were rising to values around $10^{-4}\,\mathrm{m^2/s}$. The calculation using the gradient Richardson number yielded higher values (roughly factor 4), though lower boundary values were applied and only mixing due to internal waves shear was considered.

Kocsis et al. [1998] calculated that mixing above Sill of Mainau should be enhanced by a factor of 34 compared to the lake average, which explains the discrepancy of a factor 4 from above. Keeping in mind that the Kocsis et al. measurement was done under different conditions (easterlies), and the factor 34 is a rough estimate, the ratio indicates that at least (upper estimate for Ri) $4/34 = 12\%$ of the vertical dispersive transport in MR is caused by large-scale current shear.

5. Conclusions

During the internal response the velocity profiles in western Lake Constance showed a two-layer structure, periodically changing direction as expected from measurements of the thermocline oscillation. Measurements were conducted at the measuring site MR on the Sill of Mainau in western Lake Constance. The velocity profiles corresponded well with meteorological data and the measurements at other stations.

During the zero transition of the two-layer oscillation, current profiles were measured, showing a three-layer structure, if the exciting winds exceeded 4 m/s. For weaker winds no similar feature could be found.

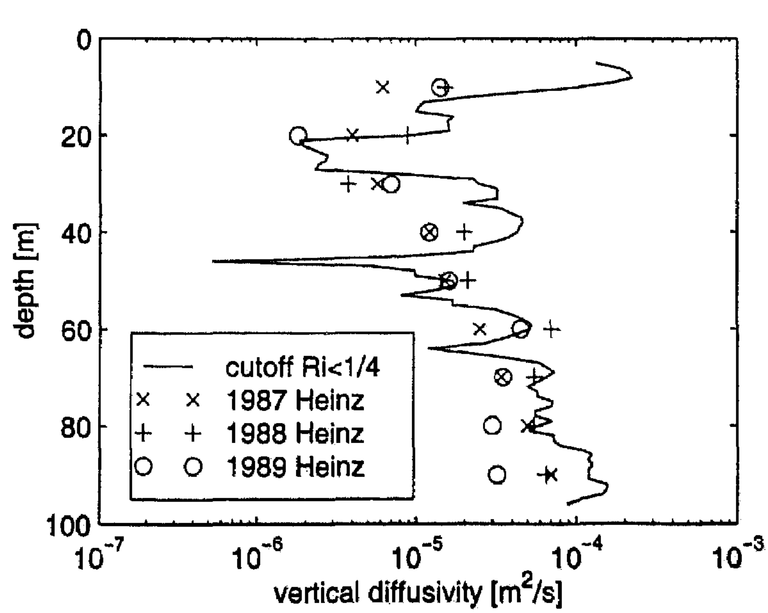

Figure 11. Vertical diffusivities versus depth; solid line: calculation from measured Richardson number in measuring site MR; symbols: measurements and calculations according to gradient flux method *Heinz et al.* [1990] for various years *Heinz* [1990].

Calculating discharges over the sill showed that some lateral structure had to be present at the beginning of the response and during periods exhibiting three-layer profiles.

Together with the simultaneously acquired temperature profiles, the velocity profiles were suited for calculating the gradient Richardson number in the epilimnion and in the thermocline and at times in the hypolimnion on a vertical length scale of about 5 m. At various depths, supercritical values were confirmed on several occasions.

In the epilimnion we confirmed supercritical Richardson numbers in the upper 10 m quite frequently, during winds especially during strong winds. With one single exception, no supercritical Richarson number was found in the thermocline. The three-layer velocity profiles caused supercritical Richardson numbers in the upper hypolimnion, 2 days after a strong wind. Later velocity profiles indicated supercritical Richardson numbers in the lower hypolimnion. In the bottom 20 m, supercritical Richardson numbers could be confirmed sporadically, with a slightly higher frequency during the internal response to strong winds.

Upper estimates for the gradient Richardson number allowed a lower estimate for the implication for vertical mixing. In conclusion, the vertical transport at MR due to large scale current shear was shown to be a factor 4 above the lake-average transport coefficient calculated by *Heinz* [1990] using the gradient flux approach through all depths. These results confirmed the importance of the sill overflow for vertical mixing in the lake during stratification.

Acknowledgments. The measurements were funded by the Deutsche Forschungsgemeinschaft, SFB 248: "Stoffhaushalt des Bodensees." Many thanks to Gerhard Heinz and Michael Schimmele for their helpfulness, while this study was carried out. They were responsible for the accompanying measurements (thermistor chain and other anchored stations) during this study. Josef Halder is thanked for his great support during field work. Karl Otto Münnich was supervising the measurements and contributed to this paper in many discussions. However, for health reasons he could not participate in the final drafting of the paper.

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B. Boehrer, UFZ-Umweltforschungszentrum Leipzig-Halle GmbH, Sektion Gewässerforschung, Brückstr. 3a, D-39114 Magdeburg, Germany (e-mail: boehrer@gm.ufz.de)
J. Ilmberger, K.O. Münnich, Institut für Umweltphysik, Im Neuenheimer Feld 229, D-69120 Heidelberg, Germany (e-mail: Johann.Ilmberger@iup.uni-heidelberg.de)

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