The Effect of the Bottom Boundary on Diapycnal Mixing in Enclosed Basins

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Abstract. Tracers released into the pelagic, stratified center of lakes, show that interior mixing—based on both mode-related internal shear and higher-frequency internal waves – is extremely weak. After horizontal spreading, the tracers "feel" the local bottom boundaries (BBL) and the basin-scale diapycnal (vertical) mixing increases by more than one order of magnitude. Balancing kinetic energy and dissipation reveals that the observed internal energetics and mixing can be explained by classical bottom friction of the basin-scale (seiches) or inertial currents alone, without relying on further processes. The assumption that the buoyancy flux is generated within the BBL and diffusively intrudes into the interior is compatible with the observed tracer residence time scale in the interior. Within enclosed water bodies of limited extent, the BBL structure is modified due to the strong periodicity (typically several hours to days) of the near-bottom currents. This Stokes' solution-like behavior (of the oscillating currents) can also be expected to be important in estuaries and ocean subbasins, where the 12 and 24-hours periodicity is relevant.

Introduction

The intensity and spatial structure of diapycnal diffusivity and upwelling are fundamental for abyssal renewal and ocean thermohaline stratification. In a steady-state ocean the stratification below the outcropping density surfaces is given by the competition between upwelling, $w\rho$, and the vertical diffusive flux, $-K{\cdot}\partial\rho/\partial z$ (*Munk*, 1966; *Munk and Wunsch*, 1998). Inverse modeling or fitting the Munk model to measured profiles of water properties provides estimates of diapycnal diffusivity in the range of the canonical value of 1 cm^2 s^{-1}. It is important to realize that such tracers are averaging processes over the entire basin and therefore capture the large-scale property distributions *Müller and Garrett*, pp. 255-263, this issue.

In contrast, by using microstructure profilers we are able to observe turbulence and mixing directly, although only in the limited volume which is sampled by the profiler. Despite the large number of microstructure observations in the 70's and 80's (*Osborn and Cox*, 1972; *Gregg and Sanford*, 1988; *Moum,* 1996), the canonical diffusivity has remained a puzzle: In fact, microstructure-based diffusivities were found to be O(0.1 cm^2 s^{-1}), which is an order of magnitude smaller than the canonical value (*Gregg*, 1987; *Davis*, 1994). This led to disputes over the potential of microstructure profilers to preferentially sense low-turbulence in the pelagic interior ocean, thus not being representative for basin-scale tracer diffusivity. In addition, the microstructure technique has been criticized for a number of methodological reasons.

This discrepancy urgently called in the late 80's for the well-known validation experiments by *Ledwell et al*. (1993, 1995, 1998, 2000). In addition, further comparison experiments were carried out in lakes (*Wüest et al*., 2000b). In contrast to the ocean, lakes have several advantages: Enclosed water bodies of manageable basin-scales allow the balancing of several water properties (such as temperature, density, or geochemical constituents). Due to their limited size, a number of processes, particularly the contribution of mixing in the bottom boundary layer (BBL), can be compared easily.

This contribution summarizes some of the peculiarities of the BBL in lakes and specifically examines its contribution to (overall) basin-scale diapycnal mixing. We show that insights into the BBL processes are important prerequisites for understanding the mechanisms of turbulent diapycnal diffusivity in lakes.

Specifics about stratification in lakes

The seasonally stratified water body in lakes (called hypolimnion) is quite different from the interior of the ocean. The stratification of the upper lacustrine thermocline is stronger and suppresses turbulence more efficiently than in the ocean. Subsequently, turbulent mixing is weak and eddy length (Thorpe) scales are short (*Lorke and Wüest*, 2002). In contrast, the stratification in deep water is usually weaker than in the ocean, leading to diapycnal diffusivities as large as the canonical value. In general, the variability of the water column stability in lakes is huge and can reach up to nine orders of magnitude: $N^2 \approx 10^{-10}$ to 10^{-1} s^{-2}.

In particular, internal waves and basin-scale motions are omnipresent and usually forced by wind (*Imberger*, 1998). Although many mechanisms of forcing are possible, it is mainly the setup of the thermocline that couples wind momentum most efficiently into the lake interior. In general, the response of the thermocline to an imposed wind stress is a complex function of geometry, stratification and the temporal dynamics of the wind forcing. For simply-shaped and small-to-medium-sized lakes, modal-type responses are usually observed (*Stevens and Imberger*, 1996; *Antenucci et al*., 2000).

For diapycnal mixing, the currents, vertical dislocations and associated shear related to internal (baroclinic) motions are most important. From the fact that about 3‰ of the wind energy flux from the atmosphere ends up in the hypolimnion (*Wüest et al*., 2000a), we can roughly estimate the mechanical energy content in the stratified interior. The level of excitation is typically ~1 J m^{-3}, corresponding to values ranging from several J m^{-2} in shallow lakes (*Gloor et al*., 2000) to almost 1000 J m^{-2} in the deepest lakes (*Ravens et al*., 2000). Given that typical energy residence time scales are days (shallow lakes) to ~1.5 months (deepest lakes), the dissipation rate ε of the internal energy is ~10^{-8} (shallow lakes) to ~10^{-10} W kg^{-1} (deepest lakes). These energy transformation rates and the modes of decay of the mechanical energy are crucial for the stratification and mixing in the interior. An important observation is that most of the internal wave energy is contained in the basin-scale waves and currents (*Imberger*, 1998). Of those, the inertial currents and the seiching are the most important energy reservoirs (*Wüest and Lorke*, 2003).

In the following sections we demonstrate how these basin-scale currents affect diapycnal mixing in enclosed water bodies of limited extent.

Dominance of BBL versus interior mixing

The basin-scale currents contain much more shear in the BBL than in the interior. This is a well-known result from the analysis of the modal structure of the internal seiches (*Fricker and Nepf*, 2000). The cause of the unexpected level of apparent turbulent diffusivity in lakes, given the very large Richardson numbers in their thermoclines, has therefore remained unexplained (*Imberger and Ivey*, 1991).

Evidence for the dominance of BBL versus interior turbulence was provided by a series of tracer experiments in lakes and deep enclosed ocean sub-basins. Carefully designed experiments using Uranin, a fluorescent artificial tracer, released into the interior of the hypolimnion of Lake Alpnach (a medium-sized Swiss lake) clearly revealed the existence of two different types of mixing regimes: slow mixing ($K_z \approx 10^{-3}$ cm^2 s^{-1}) within the stratified interior of the lake and enhanced mixing ($K_z \approx 10^{-1}$ cm^2 s^{-1}) within the BBL of a few meters height (*Goudsmit et al*., 1997). The tracer, initially released into the center of the hypolimnion - at 12 and 23 m depth—showed an extremely slow vertical spreading during the first 10 days, as long as the tracer was in the interior of the stratified water. Thereafter, the tracer cloud encountered the lake boundaries and mixing rates increased drastically (Figure 1).

The turbulent diffusivities inferred from the 2nd moment ($K = ½\ \ \sigma^2(t)/\ \ t$) of the vertical tracer distribution agreed perfectly with diffusivity estimates based on heat budgeting, which was performed as a control (*Goudsmit et al*., 1997). These experiments proved that the basin-scale buoyancy flux (left side in Eq. (1)) can be interpreted as a superposition of mixing in the interior and mixing in the BBL (right side of Eq. (1)). Expressed by the *Osborn* (1980) relation

$$K_\varepsilon(z)N^2 = \gamma_{mix}\varepsilon(z) \qquad (1)$$

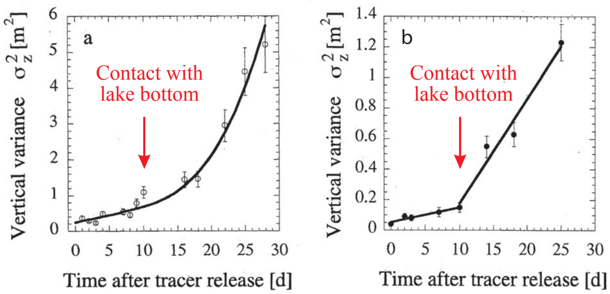

Figure 1. Two examples of vertical spreading of the dye tracer, released into the interior of the hypolimnion of Lake Alpnach in June 1995: After approximately ten days, the dye reached the BBL at the sides of the basin (arrows). The arrows indicate the drastic increase of the diffusion rate by at least a factor of ten caused by the transition from the interior to the entire basin (including the BBL). Details in *Goudsmit et al*. (1997).

this implies that ε is the sum of interior and BBL energy dissipation per kg of water at the respective depth (γ_{mix} is the mixing efficiency ~0.15).

On a much larger scale, a comparison between microstructure and "naturally" occurring tracers (i.e., already available in the lake water) in Lake Baikal revealed, again, that the basin-scale diffusivity is a superposition of contributions from the interior and from the BBL. Whereas the interior turbulence dominates in the upper thermocline (large volume per sediment area), the BBL turbulence is superior in the deeper layers (large sediment area per water volume).

Experiments by *Ledwell and Hickey* (1995) and *Ledwell and Bratkovich* (1995) in deep enclosed subbasins of limited size led to similar conclusions. Tracer inserted into the interior of the Santa Monica Basin and Santa Cruz Basin showed slow vertical spreading ($K_z \approx 0.25$ and 1 cm^2 s^{-1}, respectively) as long as the tracer resided in the interior and increased by an order of magnitude to 1.3 and 10 cm^2 s^{-1}, respectively, as the BBL became involved.

Enhanced mixing at the boundaries is manifested by the observation of well-mixed BBLs with typical heights of a few to dozens of meters in lakes (Figure 2) and many dozens of meters in the ocean (Figure 6). However, the near boundary stratification is not only determined by the generation of turbulent kinetic energy (TKE), but also by the rate of release of dissolved ions (salt) from the sediment surface (consumption of TKE). The ion flux from the sediment stems from the decomposition of organic matter, which is previously synthesized by primary production in the upper water column (*Wüest and Gloor*, 1998). For highly productive waters with low levels of kinetic energy, the redissolution of ions from the sediment can lead to strong density stratification within the BBL, which in turn suppresses turbulent mixing.

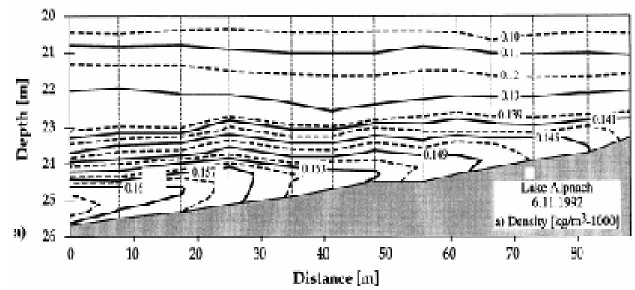

Figure 2. A snapshot of the isopycnals, determined from eleven CTD casts (collected within a few minutes) in November 1992, in the stratified water column of Lake Alpnach (max depth: 34 m). In the lower section, the BBL is thicker and unstably stratified (*Peeters et al*., 1997).

The overall effect of the two counteracting processes—turbulent mixing within the BBL and redissolution of ions—define the vertical extent of the well-mixed BBL. Conceptually, its height is related to a modified Monin-Obukhov length, which is determined by the *law-of-the-wall* generation of TKE and the buoyancy flux from the sediment ion flux.

Oscillatory boundary layers and straining-induced mixing

Usually, it is assumed that the balance between shear production and dissipation of TKE in the BBL is in a "quasi" steady-state with the forcing currents, which leads to the well-known logarithmic velocity structure and the *law-of-the-wall* dissipation profile. However, in lakes this balance is mostly unjustified for two reasons: Firstly, in low-energetic systems the turbulence can fade, especially under the mentioned ion flux from the sediment. Secondly and more importantly, due to a pronounced periodic forcing, resulting from basin-scale internal currents (seiches), the structure and dynamics of a turbulent BBL can deviate significantly from the steady-state *law-of-the-wall*. Measurements and k-ε turbulence modeling of *Lorke et al*. (2002) in Lake Alpnach—for a seiching period of 24 hours—showed that the *law-of-the-wall* scaling is restricted to the lowest 0.5 m from the sediment. Above, Stokes' solution for an oscillatory boundary layer applies, leading to a distinct maximum of the current speed at a height of a few meters above the bottom (Figure 3). In this particular case of Lake Alpnach, the divergence of TKE resulted in a phase lag between the current velocity and the turbulent dissipation of 1.5 hours. Because the phase lag and the profile distortion have been shown to also depend on the energy level (*Baumert and Radach*, 1992; *Mellor*, 2002), it is expected that the effect of the periodicity is reduced for tidal flows.

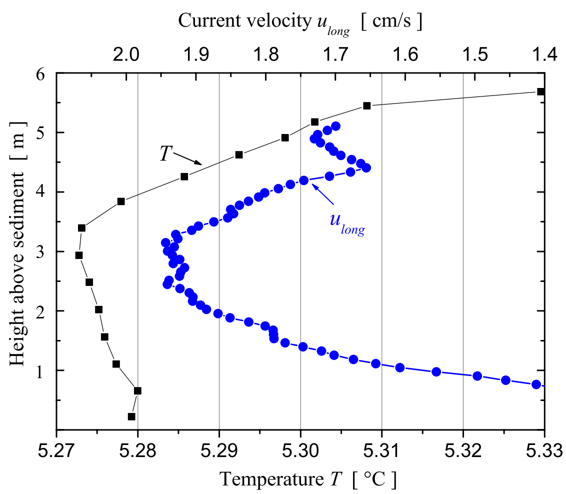

Figure 3. The Stokes' solution-type current profile within a few meters above the sediment with the characteristic maximum at 3 m. The temperature reveals the straining-induced instability in the BBL (Figure 4b). Details in *Lorke et al*. (2002).

Straining-induced convection can be an additional source of TKE, whenever the isopycnals are not parallel to the local bottom. An exemplification of this mechanism is tidal straining in the regions of freshwater influence (*Simpson*, 1997). There, the isopycnals are nearly vertical on average and the depth-varying and oscillating tidal currents create alternating stable and unstable stratification, resulting in periods of strong and weak turbulent mixing (*Rippeth et al*., 2001; Figure 4a).

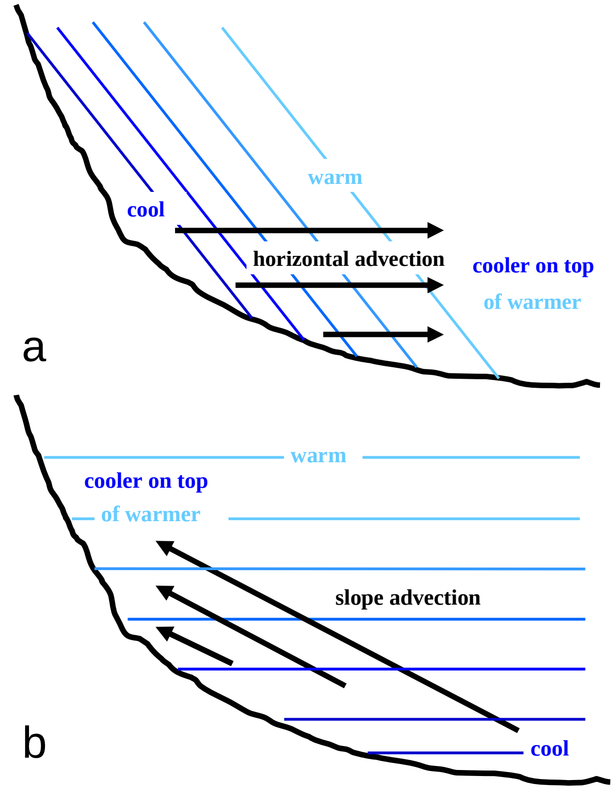

Figure 4. Schematics of two generation mechanisms for straining-induced instabilities by moving cooler water on top of warmer water: (a) Tilted isopycnals in a vertical current structure and (b) Horizontal isopycnals in an up-slope parallel current. The case (b) is the relevant mechanism for the instable BBLs of Figures 2 and 3.

An angle between the current velocity and the isopycnal surfaces can also occur above a sloping bottom (Figure 2) or in the presence of baroclinic motions (Figure 3) or by a combination of both (Figure 4). This can result in straining-induced convection analogous to tidal straining, as described above (Figure 4a). The data shown in Figures 2 and 3 were measured independently in the same lake, and both Figures unambiguously reveal the unstable stratification within the BBL. The convective layer is restricted to depth intervals with strong vertical gradients of the current speed, and hence to the vertical extent of the BBL (Figure 3). During the reversed currents, the opposite effect will stabilize the stratification and thereby suppress turbulent mixing (see *Slinn and Levine*, pp. 59-69, this issue

Interior intrusions from sloping bottoms

So far we have convincingly shown—and this is manifested by CTD profiles again and again—that turbulent mixing in stratified lakes is enhanced above the sediment (*MacIntyre et al*., 1999). As Figure 2 indicates, subsequent well-mixed BBLs build-up not only in the deepest layers, but also along the slopes. This is most relevant for the diapycnal fluxes. One of the key questions of the ‘Aha Huliko’a 2003 is therefore an intrinsic one: “How is the enhanced mixing in the BBL communicated into the interior of the stratified water, and how effective is the turbulence in the BBL for the overall basin-scale buoyancy flux?”

The buoyancy flux provides an option for relating the turbulence in the BBL to basin-scale diffusivity. Balancing the buoyancy flux, generated by intrusions of BBL water into the interior (left side of (2)), with the basin-scale buoyancy flux (right side of (2)) leads to:

$$g \cdot \langle w' \Delta\rho/\rho \rangle = K \cdot N^2 \qquad (2)$$

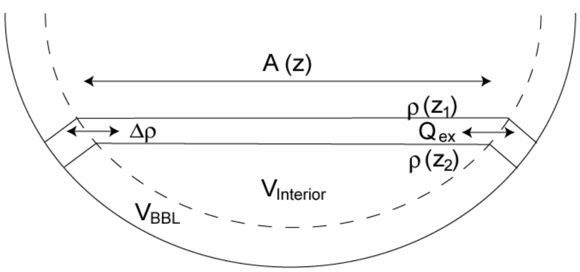

Figure 5. Schematics of the water exchange Q_{ex} between the BBL and the interior. $\Delta\rho = \rho\,(z_2) - \rho(z_1)$ is the horizontal density difference between BBL and interior water (Figure 2). $A(z)$ is the cross sectional surface at depth z.

The rate of water intrusion (expressed as upwelling velocity $w' = Q_{ex} \cdot A^{-1}$; symbols in Figure 5) is subsequently $w' = K\, N^2 \cdot (g \cdot \Delta\rho/\rho)^{-1}$, where K (~0.035 cm^2 s^{-1}) and N^2 (~$4.5 \cdot 10^{-4}$ s^{-2}) are the basin-scale diapycnal diffusivity and stability, respectively, g = 9.81 m s^{-2} and $\Delta\rho/\rho$ (~$10 \cdot 10^{-6}$; Figure 2) is the horizontal density step through the BBL. This “tertiary circulation” (*Garrett*, 1991) corresponds to an upwelling in the interior with $w' \approx 1.4$ m d^{-1} for the values given in parentheses, which are representative for this discussed case of Lake Alpnach (Figure 5; *Wüest et al*., 2000a). As the stratified volume of the lake has a mean depth of 18 m, the residence time of the interior water in the hypolimnion (V_{Interior}) is therefore 18 m / w' = 13 days. The residence time in the BBL volume is much shorter, since V_{BBL} is a small part of the entire hypolimnion. Assuming a BBL thickness of 1 m (Figure 2) yields a volume ratio of 1/18. Correspondingly, the residence time in V_{BBL} is only 13 d / 18 = 17 hours.

This time scale is indeed approximately the period of the dominant seiche (second vertical mode) and close to the residence time scale of the kinetic energy in the stratified volume (*Gloor et al*., 2000). This indicates that during basically every seiche period, the well-mixed layer of Figure 2 is newly formed during the phase of maximum currents, whereas the density difference can readjust (by releasing water into the interior) while the motion comes to a halt at the seiche’s maximum dislocation.

These estimates are quite consistent with the observations by *Gloor et al*. (2000), who concluded that the height of the BBL is not only dependent on the forcing (currents) but also on the rate of intrusions. If the BBL water were not to intrude, the BBL would become significantly higher. In fact, in the deepest layer, where there is no volume to intrude into, the thickness is indeed larger (~4 m; *Gloor et al*., 2000).

Another consistency check is provided by the TKE balance performed for the hypolimnion of Lake Alpnach. Again, comparing the total buoyancy flux of the hypolimnion (i.e., the vertical integral of Eq. (2)) with the total dissipation in the BBL allows the determination of a system mixing efficiency, by applying equation (1) in its integral form. As detailed in *Wüest et al*. (2000a), the system mixing efficiency is about 0.15 (close to the expected maximum of 0.2: *Osborn*, 1980; *Ivey and Imberger*, 1991), again indicating that the BBL water is mixed with a high effectiveness (*Garrett*, 1991). If the BBL water were to reside for long, BBL-mixing would work on water which is already well mixed and therefore we would expect reduced effectiveness (*Garrett*, 1990, 1991).

In summary: The water in the hypolimnion of Lake Alpnach resides on average for ~13 days in the interior (experiencing hardly any turbulent mixing; Figure 1) before becoming entrapped in the BBL, where it is exposed to ~300-times more intense mixing (*Goudsmit et al*., 1997). The water resides in the BBL for a seiche period (~17 hours) before it is again released into the interior. The fact that it took the tracer, after its interior release, indeed about two weeks to reach the BBL (Figure 1) is a strongly consistent argument.

Ledwell's deep ocean basins

Unfortunately, the data from the Santa Monica Basin (*Ledwell and Hickey*, 1995) and the Santa Cruz Basin (*Ledwell and Bratkovich*, 1995) are not detailed enough in the lateral dimension to perform the same estimates of the residence time scales. If we assume a density difference $\Delta\rho/\rho \approx 1 \cdot 10^{-6}$ between the interior and the well-mixed BBL (Figure 6), we obtain an approximate time scale of one month for the Santa Monica Basin ($N^2 \approx 3.7 \cdot 10^{-6}$ s^{-2}; $K_z \approx 1.3$ cm^2 s^{-1}) and one year for the Santa Cruz Basin ($N^2 \approx 1.2 \cdot 10^{-7}$ s^{-2}; $K_z \approx 10$ cm^2 s^{-1}). These values are not inconsistent with the above observations, although the temporal development is not fully resolved. *Ledwell and Hickey* (1995) conclude that a residence time of at least four days would be required in order to mix 30 m high BBLs in the Santa Monica Basin (Figure 6). Using the same buoyancy balance, we estimate a consistent value for the residence time scale of about 7 days.

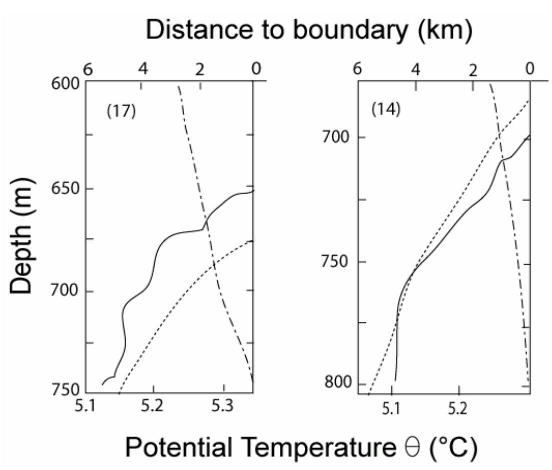

Figure 6. Two examples of the horizontal temperature structure in the Santa Monica Basin during the SF_6 tracer experiment: The profile collected in the boundary area (solid line), is compared to the profile from the interior (dotted line). Note the heights of the well-mixed layers. Details in *Ledwell and Hickey* (1995).

Conclusions

Diapycnal tracer experiments reveal that turbulence in stratified lakes mainly occurs close to the bottom boundary. The turbulence level is according to classical BBL friction and does not call for other processes, such as internal wave breaking (*Thorpe*, 2001). The observations of the tracer and kinetic energy residence time scales, as well as the observed mixing efficiency, support the hypothesis that the BBL-generated buoyancy flux is transported into the interior of the stratified water by well-mixed layer intrusions. The observations by Ledwell in deep ocean basins are not inconsistent with this conclusion.

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