F50/51 Limnophysics

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Abstract

During the practical course F50/51 the students perform measurements at Lake Willersinnweiher (an artificial lake near Ludwigshafen). Temperature, conductivity and oxygen profiles at different locations within the lake are determined. Based on these profiles the mixing behaviour of the lake is to be investigated.

In the second part of the practical course the students take water samples from the lake and a groundwater sampling site to determine the groundwater inflow into the lake. In order to examine the interaction of groundwater with the lake, the radon concentrations of the samples are measured in the hydrology lab of the Institute of Environmental Physics in Heidelberg.

Important notes:

- You won't be allowed to do this practical course if you cannot swim!

- This practical course starts on Monday at 9.00 a.m.

- In this instruction sheet all theoretical basics are included, so you do not need any further literature.

Have you seen any mistakes during reading?
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1 Motivation

We are interested in lakes for many reasons. People use lakes for relaxing after work, to go hiking around them or to just sit on a bench and enjoy nature. Furthermore, lakes are used for leisure time activities like swimming or fishing. But lakes are not only important for recreation, but also a very significant ecosystem for plants and animals.

Humans sometimes disturb the ecological balance of lakes with their various activities. To protect the ecology of lakes, it is important to know their limnological parameters. For example, if the oxygen content is too low, the fish population of the lake will possibly die off. So we need to know the oxygen content and how it is related to the mixing behaviour of the lake. One aim of this practical course is to examine the mixing behaviour of Lake Willersinnweiher near Ludwigshafen and to determine the consequences for its ecosystem. The second aim is to examine the influence of the groundwater inflow.

Figure 1.1: Lake Willersinnweiher and his neighbour gravel pit lakes, from Wollschläger [2003]

In this practical course two basic measurement techniques are applied: one limnological and one hydrological technique. To record the limnological parameters a probe is used to measure temperature, conductivity and dissolved oxygen in different depths of the lake. In the hydrological part, water samples from the lake and groundwater are taken and their radon activity concentration is determined in the lab. The measurement techniques for the radon measurement were developed at the Institute of Environmental Physics by T. Kluge [2005], C. Ebert [2007], T. Reichel [2009] and many more.

This practical course gives an overview on different fields of research and measurement techniques of the Aquatic Systems group at the Institute of Environmental Physics. The research group is subdivided into *groundwater and paleoclimate* and *physical limnology*. The members of the „groundwater and paleoclimate" group examine the noble gas concentration in combination with the age of (ground-)water and analyse the signals with regard to paleoclimate. The members of the group „physical limnology" are interested in the mixing behaviour of lakes with different sizes and shapes and their interaction with groundwater.

2 Basics

This chapter is similar to the lecture „Physics of Aquatic Systems" of Prof. Dr. W. Aeschbach-Hertig [2007/2008]. Students who took part in the lecture will recognize the theory shown in the next sections.

2.1 Limnophysics

Limnology deals with inland water and flowing water. The word limnology has its origin in the Greek word *limne*, which means lake and the Greek word *logos*, which means knowledge or lore.

Limnology is mainly a biological subject and examines the structure and function of lakes as ecosystems. To get a complete understanding of the processes in lakes, the physical and chemical relations have to be known. For example, looking at a special kind of algae which exists mainly in a depth of 3 m beneath the surface of the lake involves all three sciences. The biologist detects the algae, but the reason for this special depth could not be explained without the chemist and the physicist. One reason for this special depth is the right density of the water, where the algae can float easily and at the same time the luminosity is bright enough to allow photosynthesis.

One part of general limnology is physical limnology which deals with the physical structures and processes in lakes. One aim of physical limnology is to examine the mixing behaviour of lakes. To get information about the mixing behaviour, the stability of a water column needs to be known. The stability can be calculated with the help of water density.

2.1.1 Water density

The density ρ of a water body is defined as mass per volume and is given in $\mathrm{kg/m^3}$.

$$\rho = \frac{m}{V} \qquad \left[\frac{\mathrm{kg}}{\mathrm{m^3}}\right] \tag{2.1}$$

The density of water, like the density of other substances, depends on pressure and temperature. Additionally, the water density depends on the chemical (salt, dissolved substances, ...) and physical (isotopes) composition. The water molecule has a polar structure and is therefore a good solvent. Some special properties of water are summarized in the following table 2.1.

The density ρ of water depends on the temperature T in a special way which is very important for lakes. Due to the anomaly of density, lakes freeze from the surface to the ground. The water density in different lake depths is important for calculating the stability of the water body. The stability values provide information about the vertical mixing which is important e.g. for transporting oxygen to the lake bottom.

property	comparison with other substances	importance for the geosphere and biosphere
specific heat	higher than the specific heat of any other natural fluid	protects the environment from extreme variabilities of temperature
thermal expansion	fresh water has the highest density at a temperature above the freezing point (anomaly of density)	fresh water lakes are freezing from the surface to the ground, so fish can survive
transparency	high, because the absorption of light in the visible spectrum is small	plants in greater depth are able to do photosynthesis
heat conductivity	small	heat is transported mainly through turbulences and not through molecular diffusion

Table 2.1: Properties of water, from Aeschbach-Hertig [2007/2008]

Density of pure water

The dependency of the density of pure water on temperature is not linear and it is not possible to derive a theoretical equation. Chen and Millero [1986] give an empirical polynomial which describes the density ρ (in kg/m^3) of pure water:

$$\begin{aligned} \rho(T) \quad = \quad & 999,8395 + 6,7914 \cdot 10^{-2} \cdot T - 9,0894 \cdot 10^{-3} \cdot T^2 + 1,0171 \cdot 10^{-4} \cdot T^3 \\ & -1,2846 \cdot 10^{-6} \cdot T^4 + 1,1592 \cdot 10^{-8} \cdot T^5 - 5,0125 \cdot 10^{-11} \cdot T^6 \end{aligned} \qquad (2.2)$$

Values of temperature T have to be put into the equation in °C. This equation is applicable for a pressure p of 1013 mbar and for temperatures between 1 °C and 20 °C.

Usually the density of a substance decreases with increasing temperature because of the characteristic thermal expansion. Water has a special behaviour because the breaking up of the lattice structure of ice causes the volume to shrink, as the distance between the molecules of water is shorter than the distance between the molecules of ice. This process and the thermal expansion overlap and lead to the density function of pure water (shown in figure 2.1).

Another possibility to describe the density is given by the thermal expansion coefficient α, defined as

$$\alpha(T) = -\frac{1}{\rho}\frac{\partial\rho}{\partial T} \qquad \left[\mathrm{K}^{-1}\right] \qquad (2.3)$$

The thermal expansion coefficient has a negative sign because an increase of density means a contraction.

The thermal expansion coefficient α for pure water changes its sign at the temperature of maximal density $T_{\rho,max}$ (value is approximately 3,98 °C). For lower temperatures the thermal expansion coefficient α is negative, for higher temperatures α is positive (shown in figure 2.2).
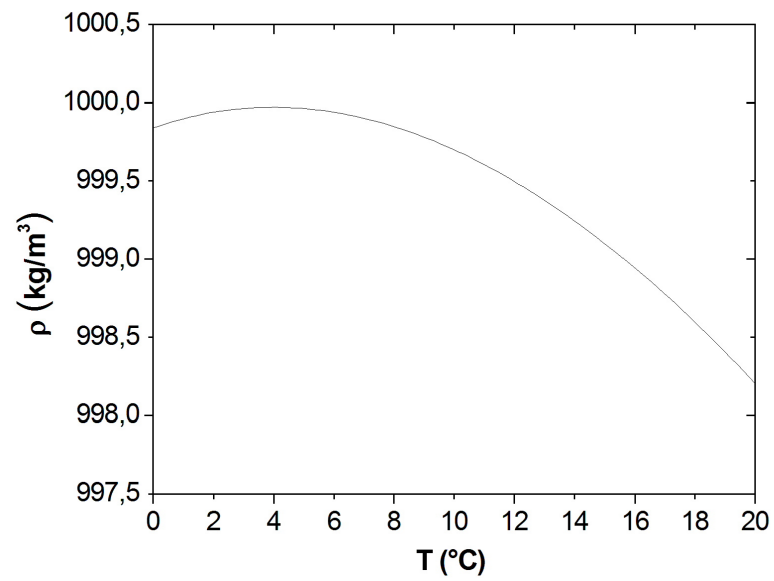
Figure 2.1: The density ρ of pure water as function of temperature T at a pressure p of 1013 mbar
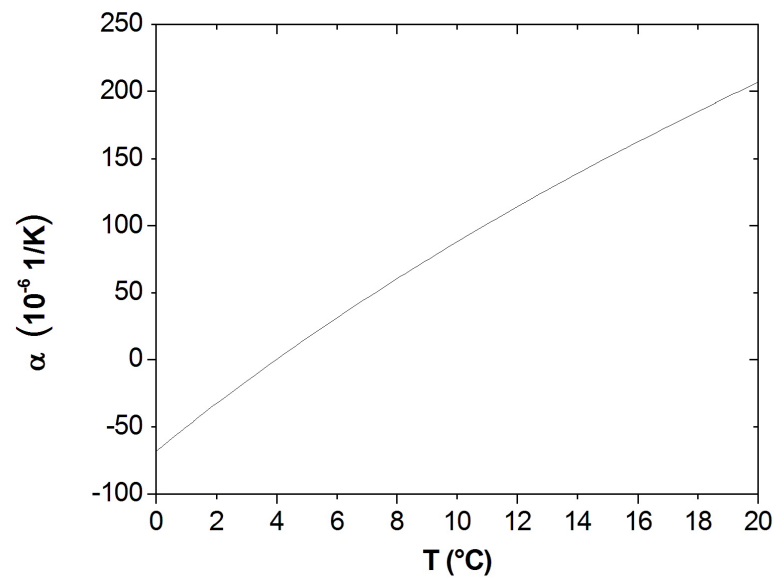

Figure 2.2: The thermal expansion coefficient α of pure water as function of temperature T at $p = 1013$ mbar, calculated as described in Chen and Millero [1986]

The density of salt water

The density of fresh water differs from the density of pure water because it additionally contains salts and dissolved gases. Usually the density is increased by dissolved salts and increased or decreased by dissolved gases. Altogether the density can be described approximately as a function of temperature and dissolved matter:

$$\rho(T, C_1, C_2, ...) = \rho(T) \cdot \left(1 + \sum_i \beta_i C_i\right) \tag{2.4}$$

Thus, the salt water density depends on the density of pure water $\rho(T)$, the different concentrations of the substances C_i and the specific contraction coefficient β_i defined as:

$$\beta_i = \frac{1}{\rho}\frac{\partial \rho}{\partial C_i} \qquad \left[\left(\frac{g}{kg}\right)^{-1}\right] \tag{2.5}$$

The specific contraction coefficient β_i describes the change of density by the change of substance concentration and is dependent on temperature. Table 2.2 shows different values of β_i at a temperature of $T = 25$ °C. Note that a negative signature indicates that the density of the water will decrease when the concentration of the substance increases.

substance	$\beta_i \left[10^{-3}\text{kg/g}\right]$
$Ca\,(HCO_3)_2$	0,813
$Fe\,(HCO_3)_2$	0,838
$NH_4\,(HCO_3)$	0,462
CO_2	0,273
NH_4	-1,250
air	-0,090
particle with $\quad\rho = 2,65$ g/cm^3	0,632
$= 1,10$ g/cm^3	0,091

Table 2.2: Influence of dissolved substances on the density of water, values from Imboden and Wüest [1995]

In general mainly salts are dissolved in lake water. Therefore, only the salts are considered and the dissolved gases and other neutral substances are neglected in the following calculations. To sum up all salts, a new parameter is introduced: the salinity S defined as the mass of all salts in a volume per total mass of the volume.

$$S = \frac{\textit{mass of all salts in a volume of water}}{\textit{total mass of the volume of water}} \qquad \left[\frac{g}{kg}\right] \tag{2.6}$$

The salinity influences the density. To get a qualitive value of the influence, the coefficient of haline contraction β_S is defined as

$$\beta_S = \frac{1}{\rho}\frac{\partial \rho}{\partial S} \qquad \left[\left(\frac{g}{kg}\right)^{-1}\right] \tag{2.7}$$

The value of β_S varies from $0{,}73 \cdot 10^{-3}$ to $1 \cdot 10^{-3}$ $(\text{g/kg})^{-1}$. The coefficient of haline concentration differs from lake to lake, but the value can be approximated if the dominating salts are known. Now it is possible to calculate the density as a function of temperature and salinity:

$$\rho(T, S) = \rho(T) \cdot (1 + \beta_S S) \tag{2.8}$$

The probe used in this practical course does not directly measure the salinity S, but the electrical conductivity κ. The electrical conductivity of a solution is defined as the reciprocal value of the specific electrical resistance ρ_S. The unit Siemens S is equal to the unit Ω^{-1}.

$$\kappa = \frac{1}{\rho_S} \qquad \left[\frac{1}{\Omega \cdot \text{m}} \right] = \left[\frac{\text{S}}{\text{m}} \right] \tag{2.9}$$

The measured in-situ conductivity is dependent on temperature and salinity. One aim of the practical course is to compare values of the salinity in different depths. Therefore, it is necessary to calculate the in-situ conductivity to a reference conductivity κ_{20} at a temperature of 20 °C. Bührer and Ambühl [1975] give an empirical polynomial for κ_{20}, applicable for lakes with calcium bicarbonate as the dominating salt.

$$\kappa_{20} = \kappa_T \cdot \left(1,72118 - 0,0541369 \cdot T + 1,14842 \cdot 10^{-3} \cdot T^2 - 1,222651 \cdot 10^{-5} \cdot T^3 \right) \tag{2.10}$$

with the electrical conductivity κ_T [μS/cm] and the measured temperature T [°C].

Now we can derive the relation between salinity S and reference conductivity κ_{20}: the salinity will increase linearly with the salt concentration if it is a highly diluted solution:

$$S = \eta \cdot \kappa_{20} \tag{2.11}$$

The value of η is approximately $0,87 \, (\text{mg/l}) \, (\mu\text{S/cm})^{-1}$ for lakes dominated by calcium bicarbonate. This means, if the conductivity increases by 1 μS/cm, the salinity will increase by $0,87$ mg/l.

Like the coefficient of haline contraction, the coefficient of the reference conductivity β_κ is defined as:

$$\beta_\kappa = \frac{1}{\rho} \frac{\partial \rho}{\partial \kappa_{20}} \qquad \left[\left(\frac{\mu\text{S}}{\text{cm}} \right)^{-1} \right] \tag{2.12}$$

After a short calculation, the equation $\beta_S = \beta_\kappa / \eta$ is derived. The value of the coefficient of the reference conductivity β_κ is calculated to $0,707 \cdot 10^{-6} \, (\mu\text{S/cm})^{-1}$ for lakes containing mainly calcium bicarbonate.

Now the relation between the salinity S and the reference conductivity κ_{20} (equation 2.11) as well as the connection between both contraction coefficients are inserted in the density equation 2.8. With this we have an expression of the density as function of the in-situ temperature and the reference conductivity which can be calculated from the in-situ conductivity measured by the probe:

$$\rho(T, \kappa_{20}) = \rho(T) \cdot (1 + \beta_\kappa \kappa_{20}) \tag{2.13}$$

2.1.2 Stratification of lakes

One of the main aspects of physical limnology is the vertical layering, called stratification. If the temperature is dominating the density, the lake is called thermally stratified. In figure 2.3 three layers are drawn in the temperature profile. The epilimnion is the upper layer with well mixed water because of the wind. In the middle layer, which is called the thermocline or metalimnion, the temperature decreases. The deepest layer is the hypolimnion.

The density is mainly influenced by temperature. Therefore, it is possible to subdivide the lake into three layers according to the temperature profile. For lakes with a high salt content this approximation is not possible. In this case the characterisation of epilimnon, metalimnion and hypolimnion has to be done with the help of the density profile.
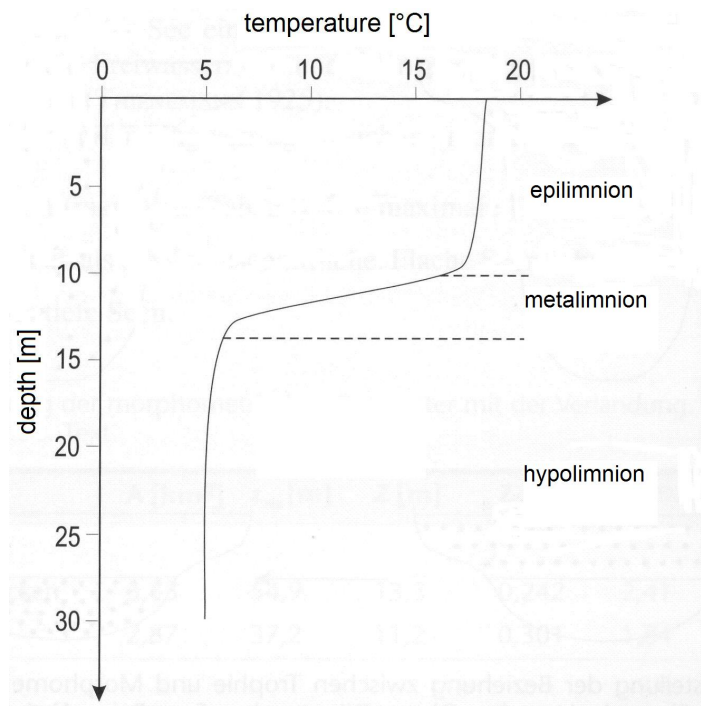

Figure 2.3: Characterisation of the different lake layers, from Schönborn [2003]

Surfaces of equal density are called isopycnals. In lakes, isopycnals are approximately horizontal. Within isopycnals no work is needed to mix water. Therefore, horizontal mixing processes are relatively fast and we consider lakes as horizontally homogeneous. Vertical mixing needs energy because work against gravity has to be done. Therefore, vertical or so-called diapycnal mixing is much slower (for further information refer to chapter 2.1.4).

The stratification of lakes changes with the seasons due to different solar radiation fluxes. Lakes get their energy mainly through the surface. Therefore, the surface temperature of a lake depends extremely on the seasonal variations of the solar energy flux. In figure 2.4 the seasonal changes of the energy flux and the surface temperature of lakes for different latitudes are shown. In our latitude the solar flux has its maximum during summer. The surface temperature reaches its maximum a little later during the late summer months. Figure 2.5 shows a typical seasonal change of stratification.
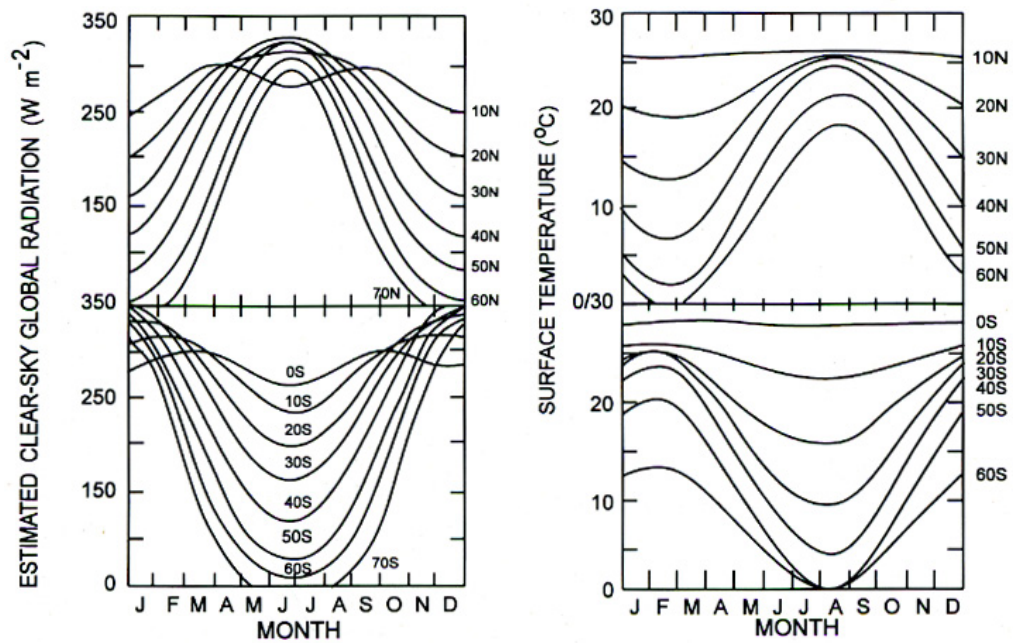
Figure 2.4: Seasonal change of solar flux and surface temperature, from Hostetler [1995]
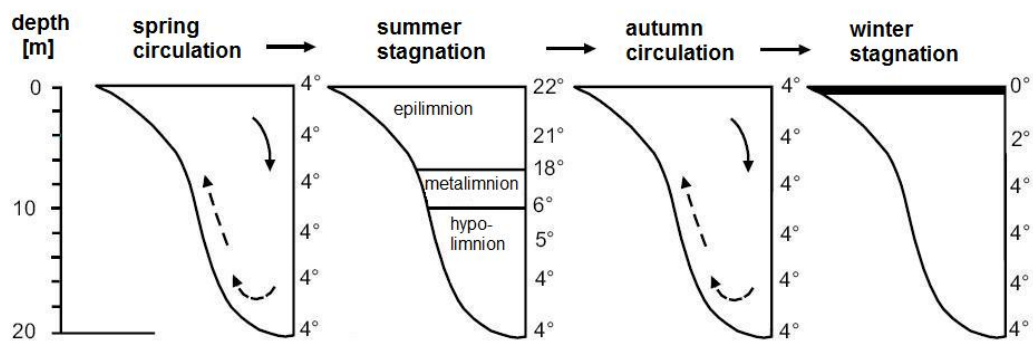

Figure 2.5: Seasonal change of stratification in a dimictic lake, from Schwoerbel and Brendelberger [2005]
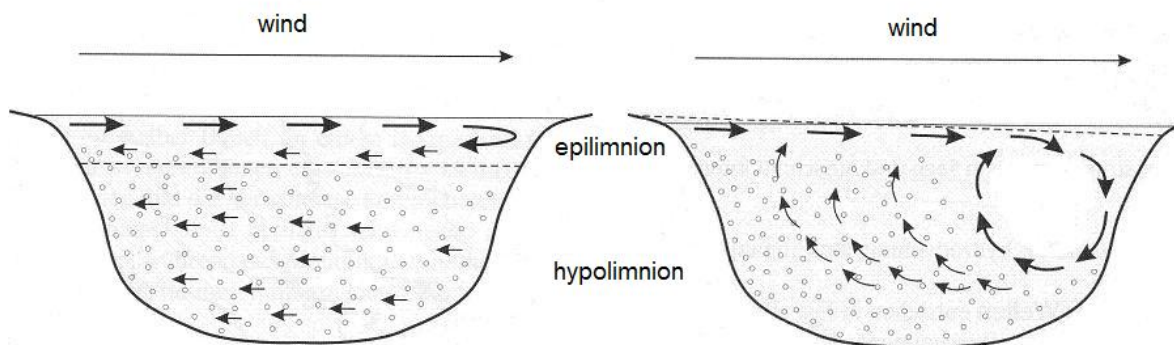

Figure 2.6: Circulation only in the epilimnion (left) and the starting circulation in the hypolimnion (right), from Schönborn [2003]

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During summer the temperature of the epilimnion increases due to the increasing solar flux. Lake water absorbs the energy of the sun, but radiation reaches only some meters down and causes no direct heating of water in the deeper part. The thermocline is the layer in which only a small part of the light arrives. Thus, the temperature in the thermocline decreases strongly. In the deeper part, the hypolimnion, lake water with the highest density is found. In deep lakes the temperature of the hypolimnion is approximately 4 °C. In this case we speak of a thermally stratified lake or the summer stagnation. Only the epilimnion is well mixed because of the wind (shown in figure 2.6 on the left side).

In autumn the solar flux is smaller than in summer. Therefore, the surface temperature decreases until it reaches approximately 4 °C. In this situation the wind is able to mix the lake completely (shown in figure 2.6). After a complete circulation the temperature of the lake is equal in every depth. This situation is called a homothermal lake.

During winter the surface temperature can decrease to values below 4 °C and the water at the surface can freeze. The ice or cold water stays above the water with a temperature of 4 °C due to the density anomaly of water (shown in figure 2.1). In this situation, called inverse stratification, the eplilimnion cannot mix with the hypolimnion without an input of energy. If the temperature at the surface decreases below the freezing point, the ice layer insulates the lake water. Therefore, the ice layer grows only slowly from the surface to the ground. This effect is very important, because it protects the fish and other animals in the lake.

In spring the ice melts and a homothermal situation is reached again because of the increasing solar flux. Now the wind causes a second complete circulation of the lake. Lakes which mix in autumn and spring are called dimictic lakes. There are other types of lakes which mix only once a year (mainly in spring).

Lakes can be classified according to their annual mixing behaviour. There are amictic, holomictic, dimictic, polymictic and meromictic lakes. Amictic lakes never circulate because they are permanently frozen or thermally stratified. Examples are lakes in the Arctic (Greenland), in the Antarctic and in the tropics. Holomictic lakes completely mix once, dimictic lakes twice and polymicitic lakes many times a year. Polymictic lakes are often relatively shallow.

Meromictic lakes do not mix completely over a long time period. In meromictic lakes a deep layer can be found, the monimolimnion, which does not take part in a circulation event for a long time. The layers above, which take part in the seasonal circulation, can be summarised as mixolimnion. This special type of stratification is often caused by a high salinity in the monimolimnion. Because of the high salinity the density of the monimolimnion is much higher than the density of the mixolimnion even if the temperature of the mixolimnion is approximately 4 °C.

In the following figures the temperature or conductivity profiles for different lake types are shown. Lake Alpnach (a part of Lake Lucerne) is a holomictic lake. From figure 2.7 we can conclude that it completely mixed in the spring of 1992.
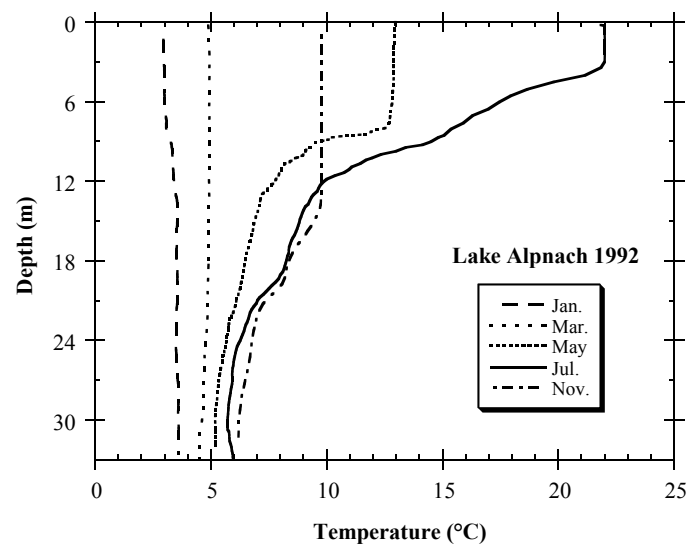
Figure 2.7: Temperature profile of Lake Alpnach, from Kipfer et al. [2002]

A meromictic lake is, for example, the artificial lake Merseburg-Ost 1b (Lake Raßnitz) in Germany. Figure 2.8 shows that the temperature below 57 m a.s.l. (above sea level) stays constant over the whole year. The temperature in the mixolimnion decreases to approximately 4 °C meaning that the temperature of the monimolimnion is higher than the temperature of the mixolimnion. This situation can only be explained by extraordinary high salinity in the monimolimnion which stops the mixing. The high salinity causing the long-time stability of the monimolimnion is shown in the right part of the figure.

Figure 2.8: Temperature and conductivity profiles of Lake Merseburg-Ost 1b, from [von Rohden and Ilmberger, 2001]

2.1.3 Stability of a water column

In the section above the vertical stratification of lakes was discussed qualitatively. Now, the Brunt-Väisäla frequency will be introduced to get a quantitative measure for the stability of a water column.
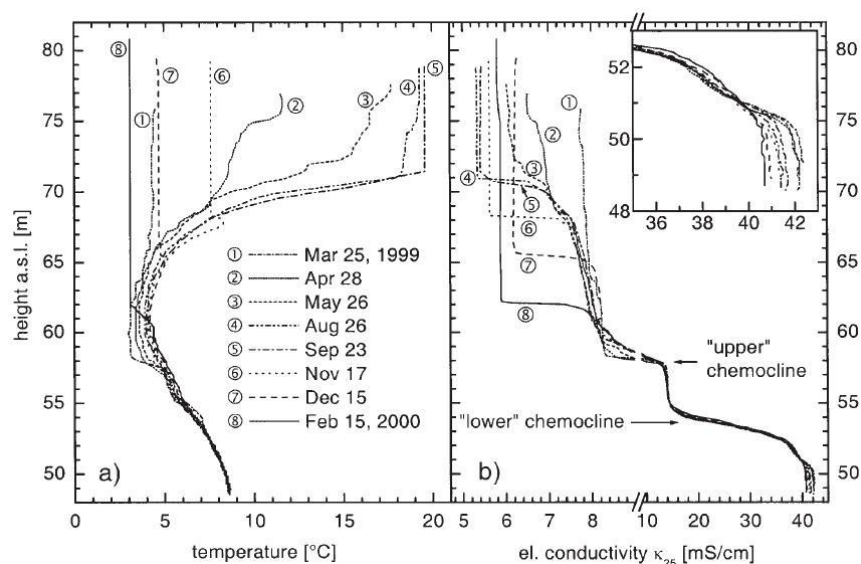
A column of water is called stable, if a water parcel, which is deflected from its equilibrium state, experiences a restoring force. Looking at a water parcel and its surrounding water the density of the water parcel ρ_P and the surrounding water $\rho_U(z_0)$ are equal at the depth z_0 (equilibrium state). If the water parcel is lifted upwards a small distance $(z - z_0)$ (shown in figure 2.9), the following forces act on the water parcel (z coordinate positive upwards):

$$
\begin{aligned}
\text{gravity:} \quad F_G &= -g\rho_P V & (2.14) \\
\text{buoyancy:} \quad F_A &= g\rho_U V & (2.15) \\
\text{force:} \quad F_{sum} &= F_G + F_A \\
F_{sum} &= -g(\rho_P - \rho_U)V & (2.16) \\
\text{acceleration:} \quad a &= \frac{F_{sum}}{\rho_P V} & (2.17)
\end{aligned}
$$
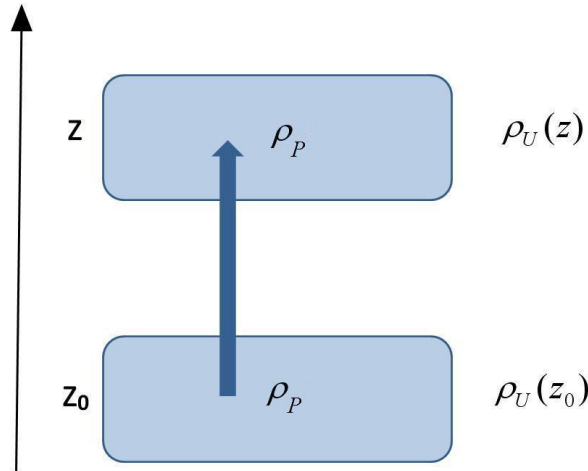
Figure 2.9: Lifted and equilibrium state of a water package

Therefore, we get the restoring force per unit of mass acting on the water parcel:

$$
\ddot{z} = -g\frac{\rho_P - \rho_U(z)}{\rho_P} \tag{2.18}
$$

The density of the surrounding water $\rho_U(z)$ at the lifted depth z is almost equal to the density of the surrounding water $\rho_U(z_0)$ at the depth z_0. Therefore, a Taylor series approximation is possible:

$$
\rho_U(z) \approx \rho_U(z_0) + \left(\frac{d\rho}{dz}\right)(z - z_0) \tag{2.19}
$$

This approximation is put into equation 2.18. Further, we consider that the density of the water parcel ρ_P is equal to the density of the surrounding water $\rho_U(z_0)$ in the equilibrium state. We derive the restoring force per unit of mass:

$$\ddot{z} = \frac{g}{\rho_P}\left(\frac{d\rho}{dz}\right)(z - z_0) \tag{2.20}$$

This is the well-known differential equation of a harmonic oscillation oscillator.

$$\ddot{z} = -N^2(z - z_0) \tag{2.21}$$

with the oscillation frequency N

$$N^2 = -\frac{g}{\rho_P}\left(\frac{d\rho}{dz}\right) \tag{2.22}$$

As discussed in chapter 2.1.1 the density of water is a function of temperature and salinity. Hence, the density gradient can be split up using equations 2.3 and 2.7.

$$\frac{d\rho}{dz} = \frac{\partial\rho}{\partial T}\frac{\partial T}{\partial z} + \frac{\partial\rho}{\partial S}\frac{\partial S}{\partial z} \tag{2.23}$$

$$= -\alpha\rho\frac{\partial T}{\partial z} + \beta\rho\frac{\partial S}{\partial z} \tag{2.24}$$

Now it is visible that the stability frequency has also two components:

$$N^2 = g\left[\alpha\frac{\partial T}{\partial z} - \beta\frac{\partial S}{\partial z}\right] \tag{2.25}$$

$$= N_T^2 + N_S^2 \tag{2.26}$$

Let's have a closer look at the derivation. So far we neglected that the temperature of the water parcel decreases as it is lifted up, because the pressure decreases. As a result of the decompression the water parcel will expand, thereby spending energy. Therefore, the temperature will decrease. The adiabatic temperature gradient $(dT/dz)_{ad}$ describes this change of temperature:

$$\left.\frac{dT}{dz}\right|_{ad} = -\frac{gT_{abs}\alpha}{c_p} = -\Gamma(T,S,p) \tag{2.27}$$

g	gravity constant
T_{abs}	in-situ temperature in Kelvin
c_p	specific heat capacity
$\Gamma(T,S,p)$	adiabatic temperature gradient
α	coefficient of thermal expansion

Therefore, it is necessary to compare two water parcels at the same reference depth. The potential temperature $\Theta(z, z_0)$ is the temperature of a water parcel from the depth z after lifting it adiabatically to the depth z_0, correcting the depth dependency of temperature:

$$\Theta(z, z_0) = T(z) - \int_z^{z_0} \Gamma\left[\Theta(z, z'), S(z), p(z')\right] dz' \tag{2.28}$$

For temperatures above 4 °C it is possible that the temperature increases a little bit with depth but the water column is still thermally stratified. Therefore, for a correct calculation of the stability frequency we have to use the potential temperature instead of the in-situ temperature. During the derivation of the Brunt-Väisäla frequency we split up the density gradient. At this point of the calculation we have to put in the potential temperature instead of the in-situ temperature:

$$\frac{d\rho}{dz} = \frac{\partial\rho}{\partial\Theta}\frac{\partial\Theta}{\partial z} + \frac{\partial\rho}{\partial S}\frac{\partial S}{\partial z} \tag{2.29}$$

The partial derivative of the potential temperature can be deduced from equation 2.28:

$$\frac{\partial\Theta}{\partial z} = \frac{\partial T}{\partial z} - \left(\frac{\partial T}{\partial z}\right)_{ad} \tag{2.30}$$

The adiabatic temperature gradient for lakes is usually small enough to be neglected. Therefore, we can use the in-situ temperature instead of the potential temperature for our lake. This approximation is not allowed in deep lakes and the ocean.

Interpretation of the Brunt-Väisäla frequency

The inverse of the Brunt-Väisäla frequency N is the oscillation period of a water package deflected by a small distance from its equilibrium state. We consider that there is no exchange with the surrounding water during the oscillation. The Brunt-Väisäla frequency squared, N^2, is a quantitative measure for the stability of a water column:

$$N^2 = \begin{cases} > 0 & \text{stable} \\ = 0 & \text{labile} \\ < 0 & \text{unstable} \end{cases}$$

For negative density gradients, i.e. if there is lighter water above heavier, the square of the Brunt-Väisäla frequency is positive and the water column is stable. For positive density gradients the value of N^2 is negative and the water column is instable.

The stratification is more stable if the density gradient is greater. Therefore, in a strongly stratified lake turbulences will calm down quickly.

The Brunt-Väisäla frequency consists of two components: N_T^2 and N_S^2. If the density gradient is dominated by temperature the stability N^2 is almost equal to N_T^2. If the density gradient is dominated by salinity the stability N^2 is almost equal to N_S^2. Usually N_T^2 dominates in the thermocline while N_S^2 characterizes stability in the hypolimnion.

In meromictic lakes the temperature component, N_T^2, may indicate instability in the monimolimnion, but the salinity is high enough so that N_S^2 is much greater than N_T^2 and finally leads to a positive N^2 which corresponds to a stratification. Because of the high stability in the salinity component, N_S^2, the monimolimnion cannot take part in the circulation. An example for this situation is shown in figure 2.10 below 60 m a.s.l.

For further information about stability take a look at Millard et al. [1990] or Wüest et al. [1996].
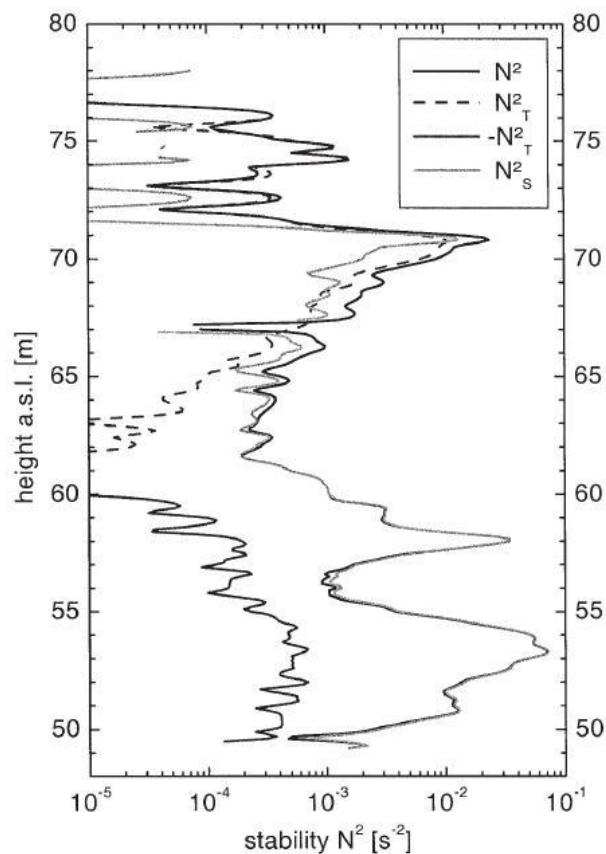
Figure 2.10: Stability of a water column N^2, and its components, for Lake Merseburg-Ost from von Rohden and Ilmberger [2001]

2.1.4 Vertical mixing behaviour

Overview

Lakes mix because of many reasons. One reason for mixing are wind-driven currents. Accelerated water with a certain velocity rubs against deeper water with low velocity. Therefore, we get a velocity gradient which causes turbulences and mixing. Another reason for mixing is an unstable water column which causes convection and turbulences.

By speaking of horizontal and vertical mixing we mean mixing along isopycnal and diapycnal surfaces. Isopycnal mixing means mixing within a density layer (density layers in lakes are almost horizontal) while diapycnal mixing takes place perpendicularly to a density layer. During isopycnal mixing we have a energy loss only due to friction. Diapycnal mixing needs energy to move water against the gravity. Therefore, isopycnal mixing coefficients are about ten or hundred times bigger than diapycnal mixing coefficients.

In a stratified lake mixing in the hypolimnion will mainly be caused by eddy diffusion and molecular diffusion. Eddy diffusion results from horizontal currents along different layers (for example sediment, isopycnals, ...). These small eddys cause further small eddys and a heat exchange is reached. Molecular diffusion needs a concentration gradient and causes also a heat exchange.

The eddy and molecular diffusion together with further mixing processes - which are described in Imboden and Wüest [1995] - are included in the mixing coefficient K_z, which is also called turbulent diffusion coefficient. There are many methods to determine the mixing coefficient. With our equipment we get the best results with the budget-gradient method.

Budget-gradient method

The budget-gradient method is based on the change of a parameter (for example temperature) below a certain depth caused by the vertical flux. We will derive the mixing coefficient by connecting the amount of heat below a layer and the vertical temperature gradient.

The budget-gradient method assumes a horizontally homogeneous lake with no vertical advection. Now we can calculate the amount of heat $W(z_0)$ in the layer between the depth z_0 and the maximum depth z_m by integration. We integrate the product of the temperature profile $T(z)$ and the cross-section area profile $A(z)$ of the lake:

$$W(z_0) = c_p \rho \int_{z_m}^{z_0} A(z)T(z)dz \tag{2.31}$$

The amount of heat in the layer changes over time. We can calculate this change of heat by differentiating the equation with respect to time. In the following step we take into account that we can exchange integration and differentiation:

$$\frac{\partial W(z_0)}{\partial t} = c_p \rho \int_{z_m}^{z_0} A(z)\frac{\partial T(z)}{\partial t}dz \tag{2.32}$$

In figure 2.11 the transport of heat through the area $A(z_0)$ during the time between temperature profiles t_1 and t_2 is shown. The amount of heat Δ transported by diffusion has increased the amount of heat and thus the temperature in the lower layer.
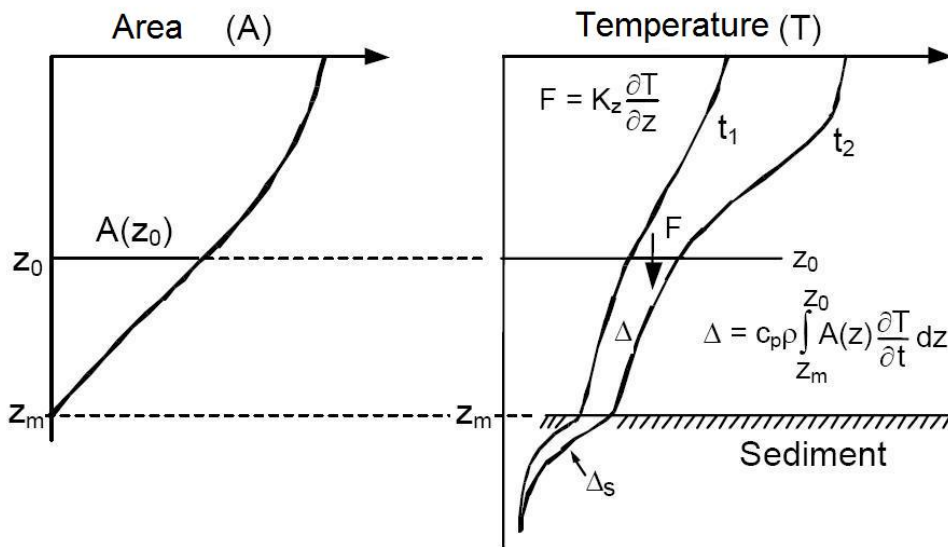

Figure 2.11: Diffusive transport of heat through the area $A(z_0)$, from the lecture Aquatic Systems I of Prof. Aeschbach-Hertig [2007/2008]

Below a certain depth we can assume a flux of heat F_{th} only caused by diffusion and not by advective transport or radiation. This vertical heat flux based on turbulent diffusion is described by Fick's first law:

$$F_{th} = -c_p \rho K_z \frac{\partial T}{\partial z} \tag{2.33}$$

F_{th} thermal flux of energy per area and time
c_p specific heat
ρ density
K_z vertical mixing coefficient
$\frac{\partial T}{\partial z}$ vertical temperature gradient

We get the additional amount of heat per time by multiplying the thermal flux F_{th} with the cross-sectional area $A(z_0)$ through which it passes. The heat flux in equation 2.33 describes the thermal flux in upward direction (z is positive upwards). We are interested in the change of heat in the volume below the depth z_0, so we consider this information of direction with the negative sign. The equation for the complete heat flux into the volume below the depth z_0 is:

$$\begin{aligned} \frac{\partial W(z_0, t)}{\partial t} &= -A(z_0) \cdot F_{th}(z_0) \\ &= A(z_0) c_p \rho K_z(z_0) \left. \frac{\partial T(z, t)}{\partial z} \right|_{z_0} \end{aligned} \tag{2.34}$$

In a first step we solve the equation for the mixing coefficient K_z. In a second step we insert equation 2.32 into the solved equation and get the following dependency:

$$K_z(z_0) = \frac{\frac{\partial W(z_0,t)}{\partial t}}{A(z_0) c_p \rho \left. \frac{\partial T(z,t)}{\partial z} \right|_{z_0}} = \frac{\int_{z_m}^{z_0} A(z) \frac{\partial T(z,t)}{\partial t} dz}{A(z_0) \left. \frac{\partial T(z,t)}{\partial z} \right|_{z_0}} \tag{2.35}$$

During this practical course the gradient is calculated from the change of two temperature profiles t_1 and t_2 which are taken at two different dates during the year. For the calculation of the mixing coefficient we therefore take the following equation:

$$K_z(z_0) = \frac{\frac{\Delta W(z_0,t)}{\Delta t}}{A(z_0) c_p \rho \overline{\left. \frac{\partial T(z,t)}{\partial z} \right|_{z_0}}} = \frac{\int_{z_m}^{z_0} A(z) \frac{\Delta T(z,t)}{\Delta t} dz}{A(z_0) \overline{\left. \frac{\partial T(z,t)}{\partial z} \right|_{z_0}}} \tag{2.36}$$

with

$$\Delta t = t_2 - t_1$$

$$\Delta W(z_0) = W(z_0, t_2) - W(z_0, t_1)$$

$$\overline{\left. \frac{\partial T(z,t)}{\partial z} \right|_{z_0}} = \frac{1}{2} \left(\left. \frac{\partial T(z,t_1)}{\partial z} \right|_{z_0} + \left. \frac{\partial T(z,t_2)}{\partial z} \right|_{z_0} \right)$$

Relationship between stratification and vertical diffusion

Stratification and vertical diffusion correlate. The relationship of the Brunt-Väisäla frequency N^2 and the mixing coefficient K_z can be described after von Rohden et al. [2007]:

$$K_z = a \cdot \left(N^2\right)^{-b} \tag{2.37}$$

Typical values of the parameters a and b for Lake Willersinnweiher (near Ludwigshafen) were determined by von Rohden et al. [2007]:

$$
\begin{aligned}
\text{a} &= (2,5 \pm 0,6) \cdot 10^{-10} \text{ m}^2/\text{s}^3 \\
\text{b} &= 0,81 \pm 0,02
\end{aligned}
$$

Equation 2.37 shows that the stratification is stronger if the vertical turbulence is weaker.

More information about this subject can be found in the papers of Quay et al. [1980], Jassby and Powell [1975] and Wüest et al. [2000].

2.1.5 Oxygen content in lakes

There are many processes which change the oxygen content of lakes. Oxygen gets into the surface water mainly by gas exchange of the surface water with the atmosphere. A further important oxygen source are green plants and cyanobacteria (blue-green algae), which perform photosynthesis. Production of oxygen by photosynthesis is dependent on light. Water absorbs light, therefore oxygen can only be produced in the upper layers. The deep hypolimnion is reached by almost no sunlight and thus it is not possible to produce oxygen there. In contrast, oxygen consuming processes are going on.

In the epilimnion the algae grow due to photosynthesis. When algae die they sink to the hypolimnion, where these organic substances are decomposed by bacteria. The decomposition is an oxygen consuming process. The oxygen in the hypolimnion which is needed for this process is mainly the oxygen of the last circulation in which the deep water exchanged oxygen with the atmosphere. The oxygen decrease starts with the stagnation of the water body.

The oxygen content in the hypolimnion is dependent on different constraints: Oxygen in the hypolimnion can only get there by mixing with the epilimnion. Therefore, the mixing type of the lake is very important for the oxygen content. A meromictic lake has no oxygen in the monimolimnion as the water of the monimolimnion has no contact to the atmosphere and the oxygen inflow by the groundwater is usually negligible. In a holomictic lake the amount of oxygen depends on the volume of the hypolimnion. A bigger hypolimnion can store more oxygen.

The strength of oxygen consumption depends on how much organic substances from the epilimnion sink to the deeper layers where they are decomposed. In a lake with only small organic production in the epilimnion it is possible that most of the organic substances are decomposed before they have reached the hypolimnion. Therefore, only little oxygen is consumed in the hypolimnion. The velocity of the decomposition of organic substances depends on temperature. At higher temperatures more organic substances are decomposed than at a lower temperature in the same time. Further, we have to consider that at higher temperatures the absolute content of dissolved oxygen is smaller than at lower temperatures (see
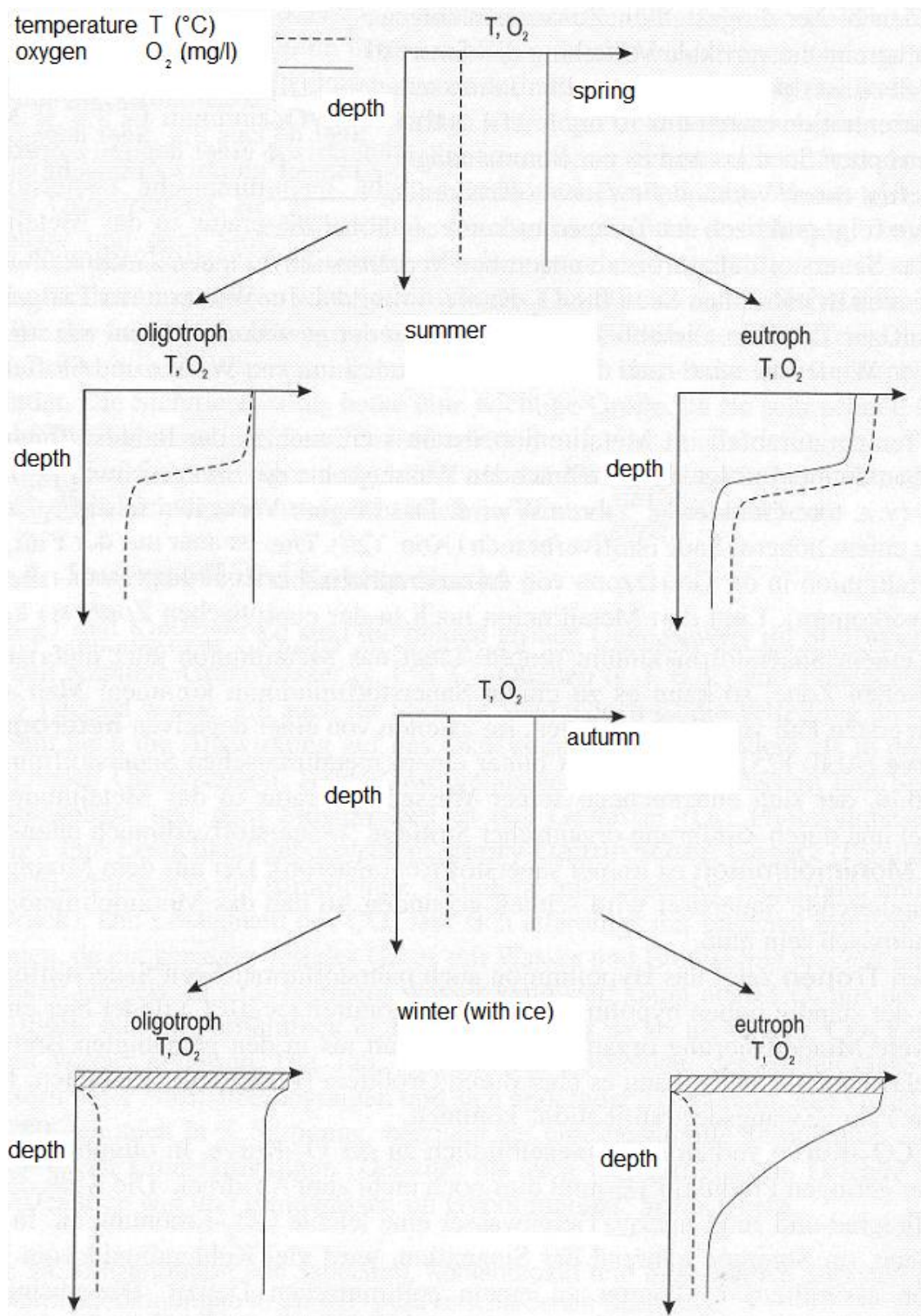
Figure 2.12: Seasonal change of the oxygen content, from Schönborn [2003]

section 2.4.3). Therefore, the oxygen content will decrease faster at higher temperatures. The oxygen profile in a lake has a characteristic seasonal change which is shown in figure 2.12.

During summer stagnation there are two typical oxygen profiles: the orthograde and the clinograde distribution of oxygen. The shape of the orthograde oxygen profile is similar to the oxygen profile after a full circulation. That means the oxygen content is nearly constant in every depth. An orthograde oxygen profile can be found in deep holomictic lakes with only small production of organic matter. Lakes with low nutrient contents and hence low productivity are called oligotrophic. The clinograde oxygen profile has a higher oxygen content in the epilimnion and a very low oxygen content in the hypolimnion. The decrease of oxygen with depth does not have to be continuous. It is possible that there are maxima and minima due to biological activity. A clinograde oxygen profile can often be found in nutrient-rich and thus highly productive lakes, which are called eutrophic lakes.

Figure 2.13 shows some examples. In part A the orthograde oxygen profile of the oligothrophic Königsee is shown from 5.7.1980. In part B and C two clinograde oxygen profiles are shown. Part B shows the oxygen profile of the deep, eutrophic Lake Biel in Switzerland from 11.10.1976 and part C shows the oxygen profile of the eutrophic Pußsee in Holstein which is sheltered from the wind (profile taken on 4.9.1989).
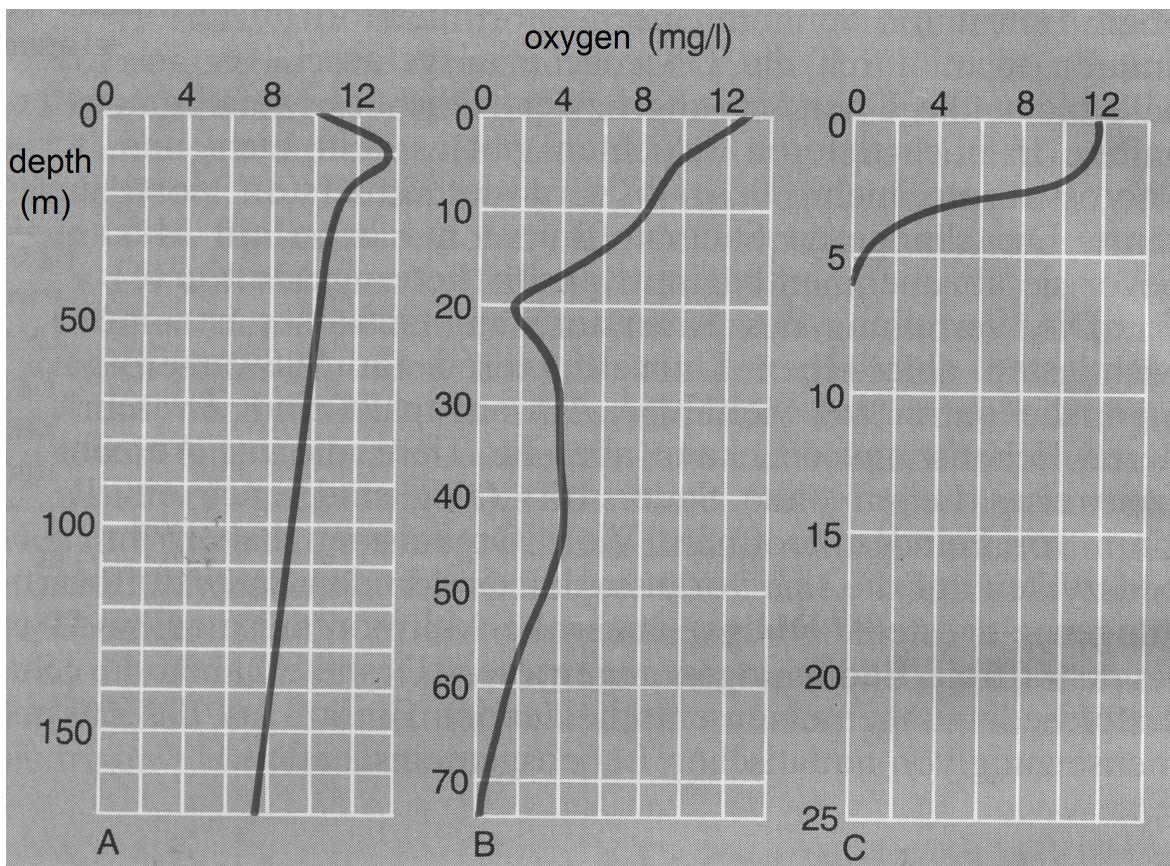

Figure 2.13: Orthograde and clinograde oxygen profiles, from Lampert and Sommer [1999]

2.2 Radioactivity and radon

2.2.1 Decay series

Radioactivity is a characteristic of unstable nuclei. They are able to transform themselves into an nucleus with a lower energetic level by emitting a characteristic radiation. In table 2.3 the three types of radiation are summarized.

type of decay	transform process	emission
α decay	$_{Z}^{A}X \rightarrow \, _{Z-2}^{A-4}Y + \, _{2}^{4}He$	helium nucleus (α-radiation)
β^{-} decay	$_{Z}^{A}X \rightarrow \, _{Z+1}^{A}Y + e^{-} + \bar{\nu}$	electron, antineutrino
γ decay	$_{Z}^{A}X^{*} \rightarrow \, _{Z}^{A}X + h \cdot \nu$	γ-radiation

Table 2.3: The three types of radiation

An unstable nucleus can decay into a stable nucleus or into another unstable one. Therefore, decay series can arise. Altogether only four different α-decay series exist, because β or γ decays do not change the mass number but the α decay decreases the mass number by four units for each decay. Today only three of the four decay series can be detected: the Thorium series, the Uranium-Radium series and the Uranium-Actinium series. The Neptunium series with a half-life of the starting isotope of 2,14 million years has already almost completely decayed.

Radon and its isotopes can be found in all three of the natural radioactive decay series, which is shown in figure 2.14.

The different radon isotopes have different half-lives. ^{222}Rn, which is sometimes called radon, from the ^{238}U decay series has a half-life of 3,82 days. The isotope ^{220}Rn, which is called thoron, from the ^{232}Th decay series has a much smaller half-life of 55,6 seconds and the isotope ^{219}Rn, which is called actinon, from the ^{235}U decay series has a even smaller half-life of 3,9 seconds.

The radioactive noble gas radon is colourless, odourless and shows a negligible reactivity with other elements. Thus, it is suitable for the use as a tracer (see section 2.4), for example, for the interaction between groundwater and lake water. The processes we look at take place in a time range of hours or days and thus we choose radon as a tracer due to its suitable half-life.

2.2.2 Decay law

The radioactive decay is a statistical process, which can be described by the radioactive decay law (equation 2.38). The number $N(t)$ of nuclei of a radioactive material at time t which have not yet decayed decreases exponentially with time:

$$N(t) = N_0 \cdot e^{-\lambda t} \tag{2.38}$$

N_0 initial number of nuclei of the radioactive material
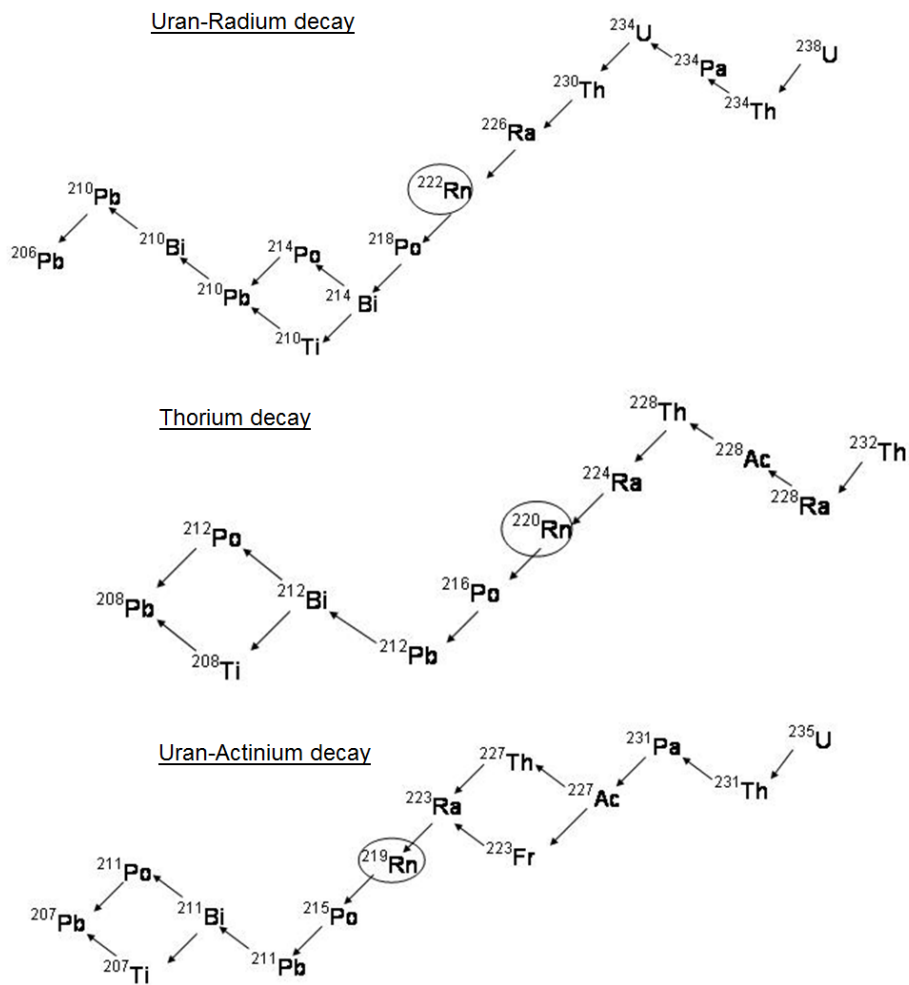λ decay constant

Figure 2.14: The three natural decay series, from Kluge [2005]

The half-life $T_{1/2}$ is the time after which the number of nuclei is halved.

$$
\begin{aligned}
N(T_{1/2}) = N_0 \cdot e^{-\lambda T_{1/2}} &= \frac{N_0}{2} \\
-\lambda T_{1/2} &= \ln\left(\frac{1}{2}\right) \\
-\lambda T_{1/2} &= -\ln(2) \\
T_{1/2} &= \frac{\ln 2}{\lambda}
\end{aligned}
\tag{2.39}
$$

Radon has a half-life of 3,82 days. Therefore, after two half-lives or approximately eight days only 25% of the original amount of radon can be measured. So, the radon measurements should be done in a short time after taking the water samples.

The activity A of a radioactive material describes the number of decays per time.

$$
A(t) = -\frac{dN(t)}{dt} = \lambda N(t)
\tag{2.40}
$$

The unit of the activity is Becquerel (Bq), Curie (Ci) or pico Curie (pCi $= 10^{-12}$ Ci). One Becquerel is equal to one decay per second. One Curie is equal to the activity of 1 g ^{226}Ra:

$$1 \text{ Ci} = 3,7 \cdot 10^{10} \text{ Bq} \tag{2.41}$$

In this practical course we will measure activity concentrations that refer to the activity of a certain water volume.

2.2.3 Secular equilibrium

We will determine the radon concentration of water by using the measurement instrument RAD7. The RAD7 is not able to measure the radon activity concentration directly, but it can measure the decay products of radon. Therefore, we derive the relationship between radon and polonium in this section.

If we look at a small decay series of a mother nucleus N_1 and a daughter nucleus N_2, the mother nucleus will decay according to the radioactive law (differential version):

$$-\frac{dN_1(t)}{dt} = \lambda_1 N_1(t) \tag{2.42}$$

The amount of daughter nuclei is dependent on the amount of recently decayed mother nuclei and is also dependent on the decay of daughter nuclei itself. The change of the amount of daughter nuclei can be described in the following way:

$$dN_2(t) = \lambda_1 N_1(t) dt - \lambda_2 N_2(t) dt \tag{2.43}$$

If we now insert the dependency between the number of decays N and the activity A, as described in equation 2.40, we get the activity of the daughter nucleus A_2:

$$\frac{dA_2}{dt} = \lambda_2 (A_1 - A_2) \tag{2.44}$$

This differential equation is solved in Wilkening [1990]. By taking the starting conditions

$$A_1(0) = A_1^0 \tag{2.45}$$
$$A_2(0) = 0 \tag{2.46}$$

into consideration the solution is

$$A_2(t) = \frac{\lambda_2}{\lambda_2 - \lambda_1} \cdot A_1^0 \cdot \left(e^{-\lambda_1 t} - e^{-\lambda_2 t} \right) \tag{2.47}$$

Figure 2.15 shows the activity of the mother nucleus Radon-222 ($T_{1/2} = 5500,8$ min) and the daughter nucleus Polonium-218 ($T_{1/2} = 3,05$ min). We infer that after a time t of approximately 20 minutes the equilibrium state is reached and that the activity of the daughter nucleus is nearly constant. This equilibrium is called secular equilibrium.

The equilibrium state can be explained mathematically by the different half-lives. The half-life T_1 of radon is much bigger than the half life T_2 of polonium. Therefore, the decay constant of radon is much smaller than the decay constant of polonium ($\lambda_1 << \lambda_2$) and we can make the following approximation:
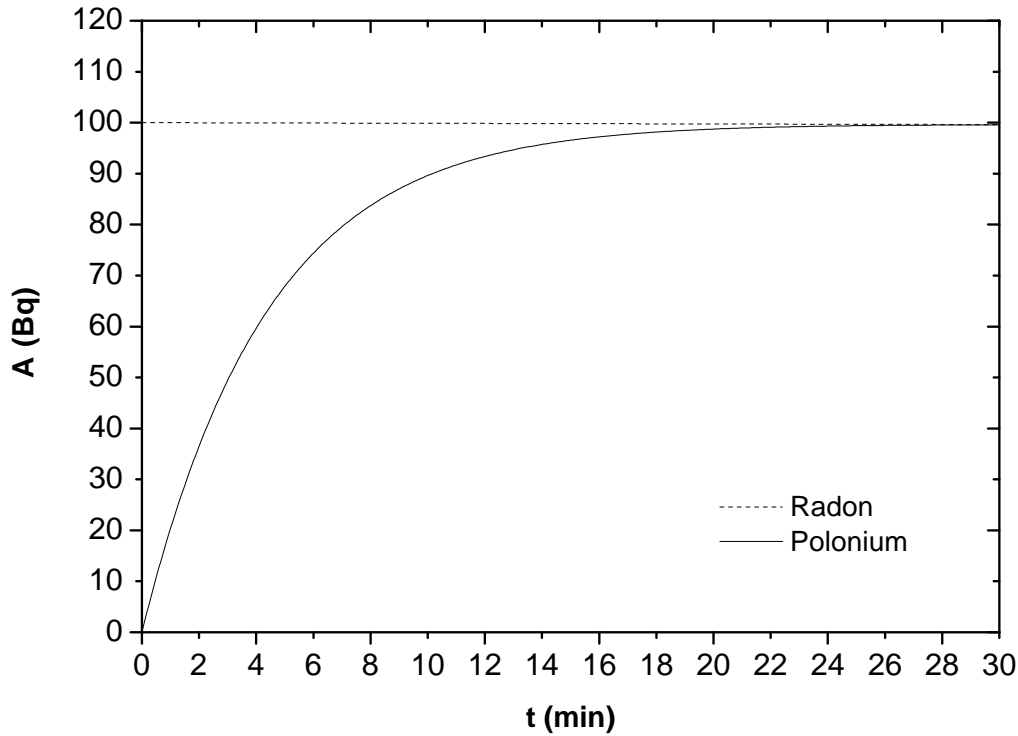
$$\lambda_2 - \lambda_1 \approx \lambda_2 \tag{2.48}$$

Figure 2.15: Secular equilibrium between the activity A of ^{222}Rn and ^{218}Po in dependence of the time t

After a period of time t, which is much longer than the half-life T_2, the term $e^{-\lambda_2 t}$ in equation 2.47 is almost zero. With this approximations and the activity of the mother nucleus $A_1(t) = A_1^0 \cdot e^{-\lambda_1 t}$ we derive mathematically the secular equilibrium:

$$A_2(t) = A_1(t) \tag{2.49}$$

In the case of a much bigger half-life of the mother nucleus than of the daughter nucleus we will always see that the activity of both radioactive materials will be equal after a suitable time t. The secular equilibrium is important for the calculation of the activity concentrations of radon with our measurement instrument RAD7.

We reach the secular equilibrium between ^{222}Rn and ^{218}Po after 20 minutes. The secular equilibrium between ^{222}Rn and ^{214}Po (see figure 2.14) is reached after three hours. This second secular equilibrium is considered for very exact measurements. During the practical course we will mainly consider the equilibrium between ^{222}Rn and ^{218}Po.

2.3 Groundwater

Groundwater is subsurface water, which flows above or between a watertight aquitard in the ground. Figure 2.16 shows the different zones of groundwater.
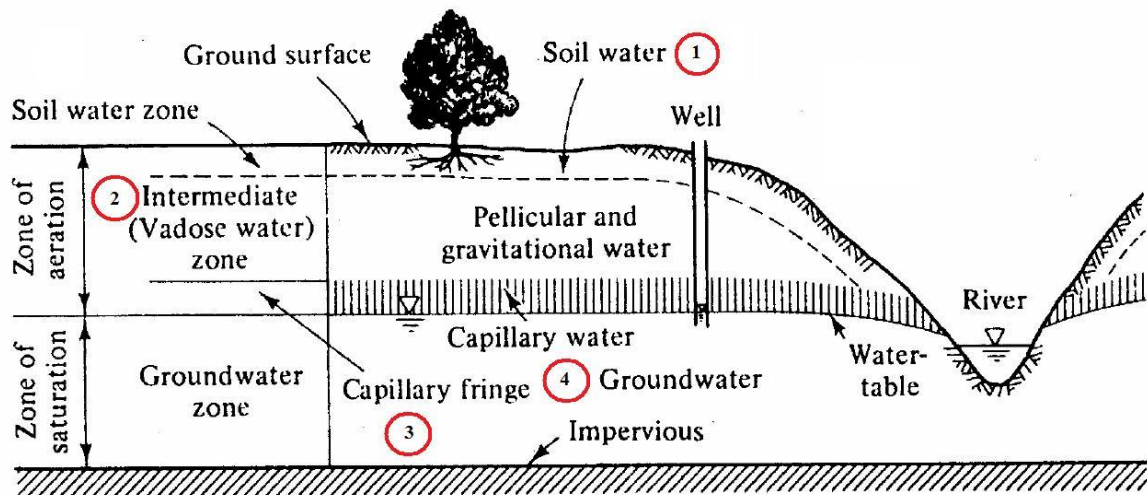

Figure 2.16: Groundwater zones, from Bear [1979]

The soil water zone (1) is a zone near the surface in which the roots of the plants hold the water. The water content of the soil water zone is dependent on precipitation. Below the soil water zone the intermediate or vadose zone (2) is located. There, water is kept by capillary forces. However, the vadose zone is not saturated and we can find air embedded in pore spaces. At the lower part of the vadose zone the capillary fringe (3) is located. Water of the capillary fringe rises from the saturated groundwater zone (4) due to capillary forces. The groundwater level is the depth of the saturated groundwater zone in which the hydrostatic pressure equals the atmospherical pressure.

Groundwater flow occurs in aquifers which are a system of connected pores, so that water can flow slowly through it. Good aquifers are e.g. sediments of gravel, sand and sandstone.

2.4 Radon as a Tracer

In this practical course we will investigate in which depth groundwater stratifies into the lake water. This is done with the help of a tracer. A tracer is a substance or a characteristic property which marks a water body or water mass. We are therefore able to detect a certain water mass. The group of tracers can be divided into natural and artificial tracers. Natural tracers are, for example, the isotopic composition of the water itself, the amount of heat respectively the temperature of a water mass or - as in our case - the radon concentration of the water. Artificial tracers like SF_6 have to be introduced deliberately into the system.

Radon is a noble gas with a very low chemical reactivity under natural circumstances. Further, there is a big difference in the concentrations of radon in lake and groundwater. Thus, radon is a good tracer for the inflow of groundwater into lakes. Temperature is not such a good tracer, because the as the temperature difference is not as large and temperature influences density and hence the depth of interstratifications of the inflowing groundwater.

2.4.1 How does radon get into the groundwater?

Radium is the mother nucleus of radon and is abundant in every type of sediment or rock in different concentrations. Radium decays to radon inside the ground or sediment and leaves the sediment grains due to the recoil from the decay (conservation of linear momentum) (see Dehnert et al. [1999]). Another possibility is that radon diffuses through the sediment into groundwater, where it accumulates.

In a homogeneous sediment a characteristic radon activity concentration can be calculated. Sandler [2000] shows that the value of the characteristic radon activity concentration for a sand and gravel aquifer is approximately 9 000 Bq/m^3.

2.4.2 How does radon get into lake water?

There are different origins of radon in lake water. Inflows and precipitation as well as the surface water are in contact with the atmosphere, thus their radon concentration is determined by the solubility equilibrium (see below). Since the atmosphere has no radon sources but only the sink due to decay, its radon content and thus that of surface water is very low. A further origin of radon in lakes is diffusion from the sediments. Another source is the decay of dissolved radium in the lake water. And finally radon in lake water originates from groundwater, if there is an inflow of groundwater into the lake.

Altogether the radon concentration in lake water is much lower than in groundwater due to firstly the exchange of the surface water with the atmosphere and secondly the quite fast decay.

2.4.3 Henry's law

We take a look at a system of two phases, for example air and water. In this system we add a gas, in our case radon. A part of this gas will dissolve into the water and the other part will stay in the air. If we wait a sufficient period of time we will see that the gas reaches a dynamical equilibrium state, which means that the flux of radon from air to water equals the flux the other way round.

The amount of gas dissolved in the water phase depends on the solubility of the specific gas and other factors. The dimensionless Ostwald solubility, a, describes the ratio of the equilibrium concentration of the gas in the fluid to the concentration in the gas phase:

$$a = \frac{C_{fl}}{C_g} \tag{2.50}$$

a Ostwald solubility
C_{fl} activity concentration in the fluid phase
C_g activity concentration in the gas phase

The solubility of radon is dependent on temperature. This dependency was found empirically by Weigel [1978]:

$$a(T) = 0,105 + 0,405 \cdot e^{-0,0502 \cdot T} \tag{2.51}$$

In equation 2.51 the temperature, T, has to be put in in °C. For example, if we calculate the solubility at a temperature of 20°C we get a value of 0,25. That means that at 20°C
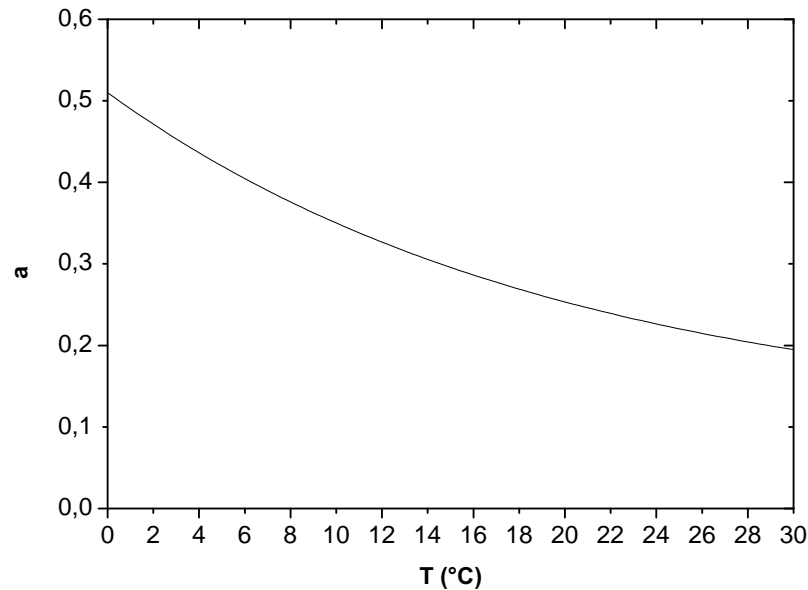
Figure 2.17: Solubility a of radon dependent on temperature T

the concentration in the gas phase is four times higher than in the fluid phase at equilibrium conditions. Figure 2.17 shows the dependency of the solubility on temperature.

In the equilibrium state the gas concentration in the water phase is dependent on its concentration in the gas phase C_g (see equation 2.50). The concentration of the gas phase is further dependent on the pressure as described in the ideal gas law:

$$C_g = \frac{n}{V} = \frac{p}{RT} \qquad (2.52)$$

C_g concentration in the gas phase in mol/liter
n   number of mols of the substance
V   volume of the gas in liter
p   pressure of radon in atm
R   ideal constant for gas (R = 0,08206 atm liter mol^{-1} K^{-1})
T   temperature in K

Henry's law states the proportionality of the gas concentrations in a liquid and a gas phase in equilibrium, as expressed by equation 2.50. Often, it is expressed as a relationship between the concentration of the dissolved gas in the fluid C_{fl} and the partial pressure in the gas phase p. We can derive Henry's law in that form from equations 2.50 and 2.52. There are different possibilities to write Henry's law. The first and commonly used possibility is with the help of the Henry coefficient $K_H = RT/a$:

$$p = C_{fl} \cdot K_H \qquad (2.53)$$

The second possibility uses the solubility coefficient $K_S = a/(RT)$:

$$C_{fl} = K_S \cdot p \qquad (2.54)$$

31

The Henry coefficient and the solubility coefficient are dependent on temperature, but not on pressure. Further, the solubility of radon is dependent on the salinity. Usually the solubility decreases with increasing salinity.

2.5 Lake Willersinnweiher

Lake Willersinnweiher is one of four artificial (dredging) lakes, located between Friesenheim and Oppau near Ludwigshafen in the upper Rhine Valley. It was created in the beginning of the 1930s by the BASF. The gravel pit reached to the upper groundwater layer (which reaches down to a depth of 25 m in the Rhine Valley). Therefore, the groundwater is connected to the lake and filled the gravel pit up with water. Soon the newly formed lake was used for swimming. Lake Willersinnweiher has no surface inflows or outflows and is only fed by groundwater and rain. Groundwater in the surroundings of the lake flows from south-west to north-east with a relatively slow velocity of $6,1 \cdot 10^{-3}$ m/d to 0,45 m/d. The values are taken from the model of Wollschläger [2003].

During autumn 1975 the lake suffered from a oxygen depletion which required an emergency aeration. After this emergency aeration the morphology of the lake was reshaped to improve the lake's ventilation. Before Lake Willersinnweiher was reshaped it was divided into two parts by an underwater barrier. The height of this barrier was decreased to enable a deep circulation between both parts. Furthermore, the mud of the sediment was dug out and the lake reached its present maximum depth of 20 m. In 1989 Lake Willersinnweiher got its official permission to be used as a swimming lake. As a consequence the shore of the lake was reshaped with some small isles. In figure 2.18 the recent morphology of Lake Willersinnweiher is shown.
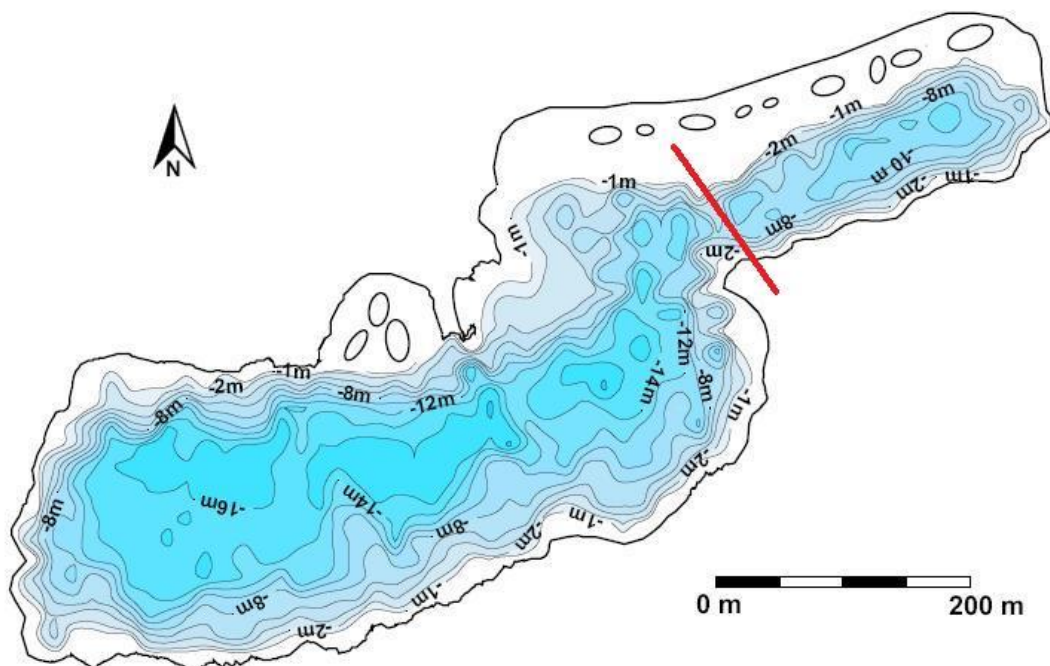

Figure 2.18: The morphology of Lake Willersinnweiher. The barrier is marked by a red line. From Wollschläger [2003]

The lake bottom is irregularly shaped with a very steep shore. The shape of the shore is almost a straight line. The red line in the figure marks the 8 m depth barrier which impedes the deep water exchange.

In table 2.4 some important parameters of Lake Willersinnweiher are summarized from the dissertations of Sandler [2000], Schmid [2002] and Wollschläger [2003]. Lake Willersinnweiher is a hard water lake with lots of calcium. Thus, the equations derived in chapter 2.1.1 are valid for Lake Willersinnweiher.

geographical parameters		typical amounts of ions in the lake water	
volume	$1,3 \cdot 10^6$ m^3	sulfate	$2,4$ mmol/l
surface area	$16 - 17$ ha	bicarbonate	$2,3$ mmol/l
mean depth	$7,7$ m	chloride	$2,2$ mmol/l
maximum depth	20 m	sodium	$1,9$ mmol/l
maximum length	850 m	calcium	$2,5$ mmol/l
maximum width	325 m	nitrate	30 μmol/l

Table 2.4: Parameters of Lake Willersinnweiher summarized from Sandler [2000], Schmid [2002] and Wollschläger [2003]

2.5.1 Mixing behaviour in the last years

Lake Willersinnweiher has been examined by the Institute of Environmental Physics for several years. Therefore, we know that the lake is a holomictic or dimictic lake. The dimictic behaviour only occurs if the ice sheet stays over a longer period or the temperatures are very low over a longer period. For some time the smaller part of the lake showed a specific behaviour characterized by an increasing concentration of salts towards the ground. Furthermore, we have detected that the circulation period gets shorter. This could be caused by a faster increase of the air temperatures in spring which stops the circulation.

2.5.2 Oxygen content in the lake

There is a lot of agriculture around Lake Willersinnweiher that causes a high inflow of fertilizers into the groundwater. As a consequence Lake Willersinnweiher is an eutrophic lake with a high biological productivity in the epilimnion. This is visible in spring when strong algal blooms occur. These algae produce a lot of organic matter which sinks to the hypolimnion and is decomposed, inducing high oxygen consumption. Therefore, the oxygen content in the deeper layers is very low during times of stagnation (values of 1,4 mg/l).

3 Measuring instruments and techniques

3.1 CTD probe

3.1.1 Design and functionality of the probe

The HYDROLAB probe is a portable sensor capable to measure in-situ temperature, conductivity, depth and oxygen concentration. Such sensors are usually called CTD probes: C for *conductivity*, T for *temperature* and D for *depth*.
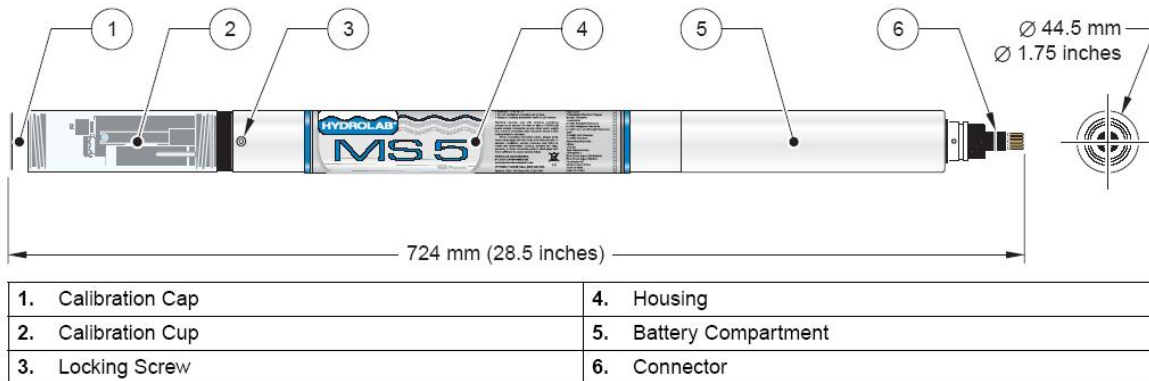

Figure 3.1: Structure of the CTD probe, from Hydrolab [2005]

The probe measures temperature with the help of a 30 kΩ variable resistance thermistor, which is able to measure temperature with an accuracy of 0,1 °C and a resolution of 0,01 °C in the range of -5 °C to 50 °C. The conductivity sensor consists of four graphite electrodes in an open cell and is able to measure conductivity in a range of 0 to 100 mS/cm with an accuracy of 1 μS/cm and a resolution of 0,1 μS/cm. Depth information is collected with a pressure sensor, that is able to measure depth in the range of 0 m to 100 m with an accuracy of 5 cm and a resolution of 1 cm. The oxygen sensor consists of a measuring chamber (a so called Clark Cell) and a circulator which helps to get a sufficient sample flow across the membrane of the chamber. The sensor measures oxygen by electrochemical reduction of oxygen diffusing through the selective membrane of the chamber. The oxygen sensor is able to measure with an accuracy of 0,2 mg/l in the range of 0 to 20 mg/l and with an accuracy of 0,6 mg/l in the range of 20 mg/l to 50 mg/l. Both ranges have a resolution of 0,01 mg/l.

3.1.2 A short introduction to the software of the probe

The CTD probe is controlled by the software Hydras 3 LT. Therefore, connect the probe to the laptop and start the software. Usually the software detects the probe in the main menu. If this is not the case, plug the USB wire out and in again and click on the button `Re-Scan for Sondes`.
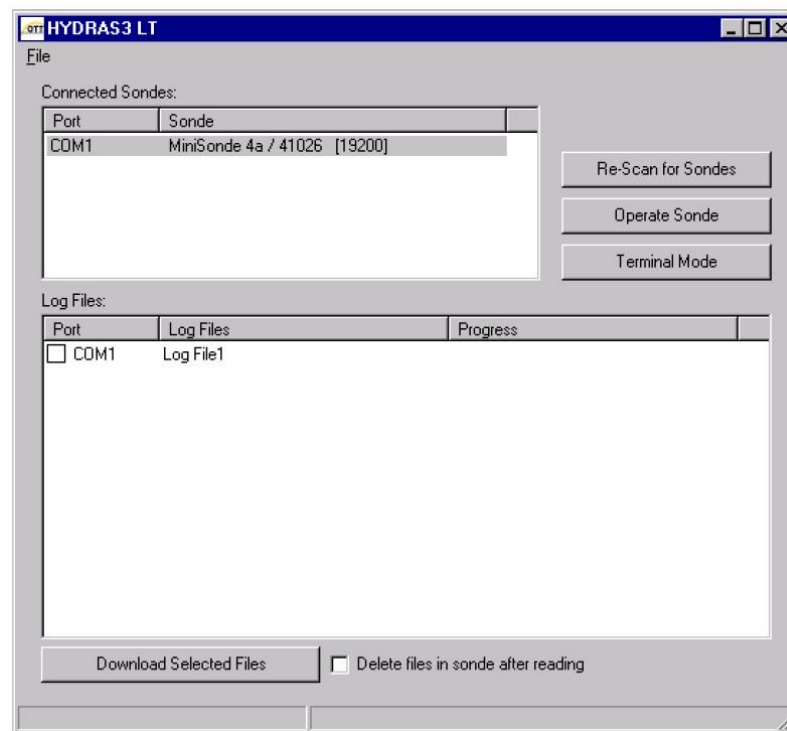
Figure 3.2: Starting window of Hydras3LT software, from Hydras3LT [2004]

As a first step we have to tell the software which parameters we like to store. This is possible in the window `Operate Sonde`. The `Operate Sonde` menu is divided into several tabs. In the tab **Online Monitoring** we first choose the Monitoring Mode (1) „time series" which means that after each time unit (choose the time unit in the Monitoring Interval (2)) the probe will perform a measurement and will store it. Figure 3.3 shows that after every 5 seconds one measurement is stored. In the window Parameters (3) we choose the parameters we want to store. For this practical course we need temperature (Temp in °C), conductivity (SpCond in μS/cm), dissolved oxygen (DO% in % and DO in mg/l), depth (Dep100 in m) and the circulator.

If you want to change the order of the parameters use the up and down arrows (4). There are different possibilities to present the data (5): In a graphic (`New Graph`) or in a table (`New Table`). At the end of one whole measurement the data should be exported to an excel file and a text file. Please store the files with the following name: year month date and measuring site. For example, if a data series was taken on September the 15th 2008 at measuring site A you store the file with the name 080915A

You can observe the data during measurement (figure 3.4) and are able to see changes in the parameters accordingly to the time interval you have chosen. If you want to change the shape of the graphic just right-click.

For further informations to the Hydrolab probe look into the manual which can be found on the laptop, in the lab or ask your supervisor.
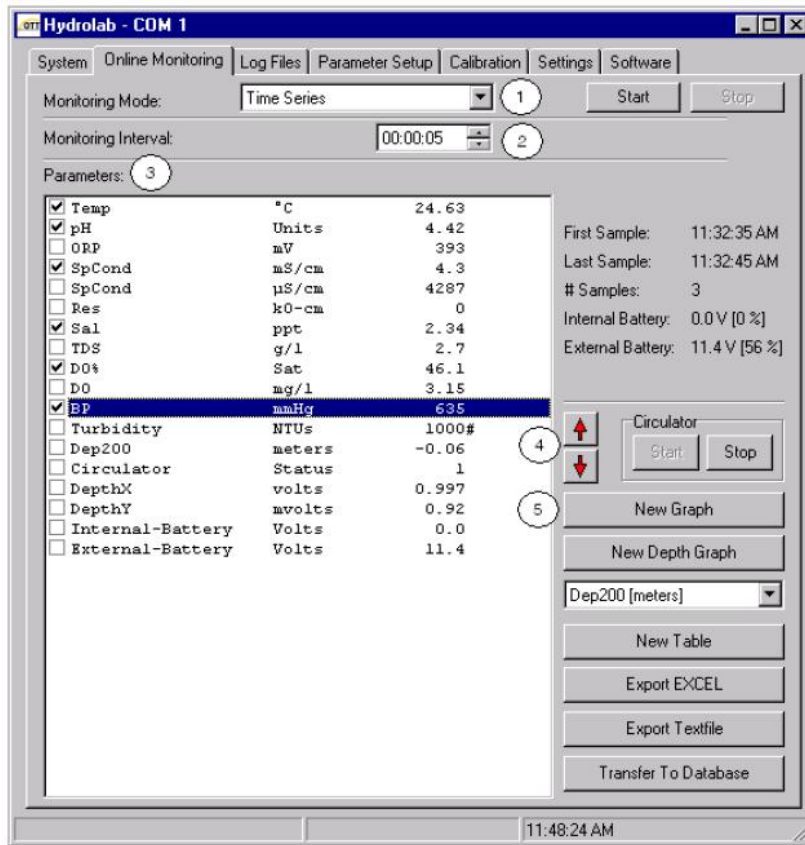
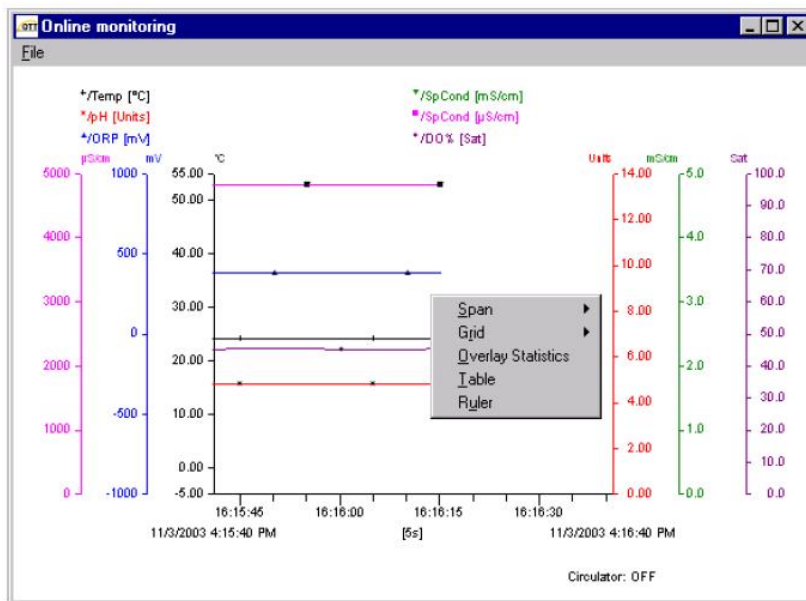Figure 3.3: The *Online Monitoring* tab in the *Operate Sonde* menu, from Hydras3LT [2004]

Figure 3.4: Window „Graph" during measurement, from Hydras3LT [2004]

3.2 Radon measurement instrument RAD7

3.2.1 How the RAD7 works

In the interior of the measurement instrument RAD7 from Durridge we find a hemisphere with a silicon solid-state detector. A representation of the measurement chamber with the detector is shown in figure 3.5.
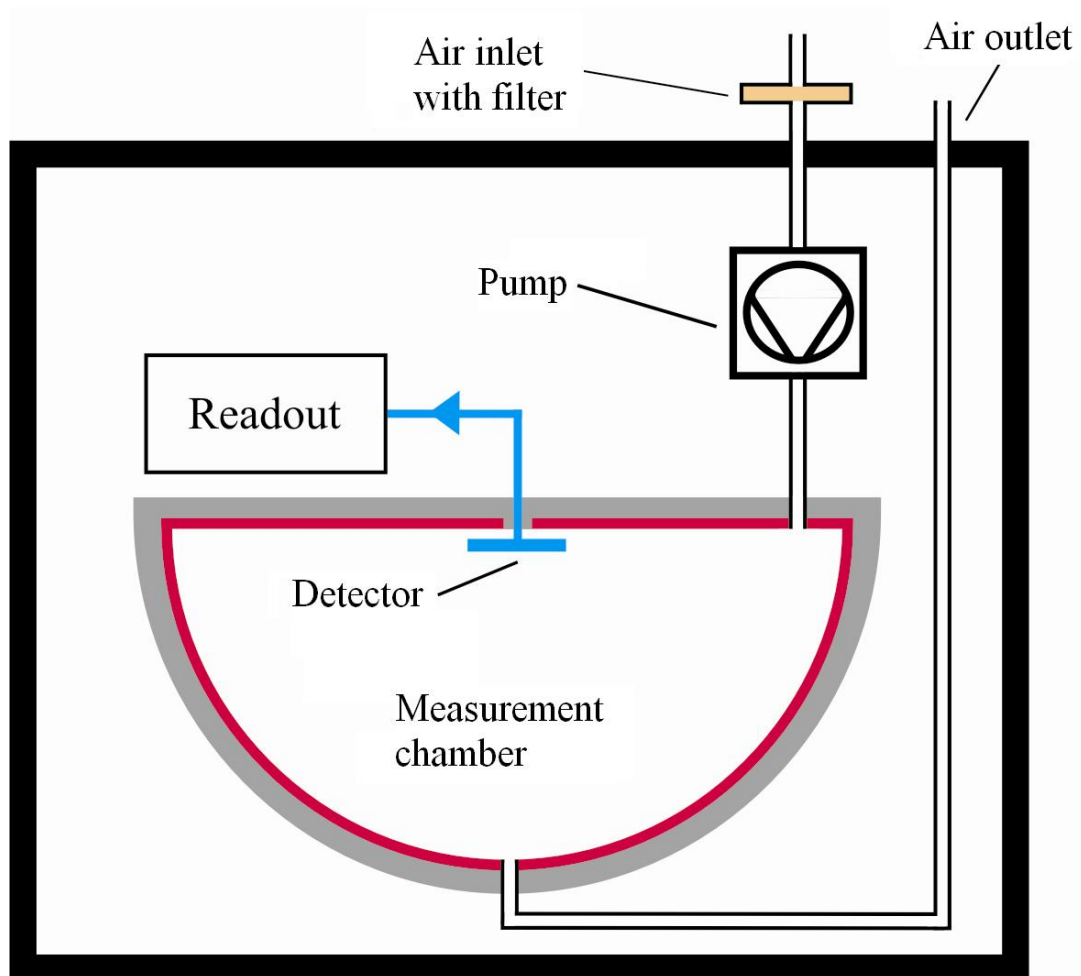
Figure 3.5: Measurement chamber of the RAD7, from Reichel [2009]

Through the filter the sample air is sucked in by the pump and reaches the detector chamber. There a high voltage of 2000 to 2500 V between the detector and the hemisphere accelerates the positively ionized particles towards the detector. If a radon nucleus decays in the chamber into a positively ionized polonium-218 this particle will be accelerated towards the detector. On the surface of the detector the short lived ^{218}Po decays and the α radiation with a characteristic energy is emitted to the detector. The detector produces a signal with 50 per cent probability. This signal is intensified electronically and transformed into a digital signal. The microprocessor stores the energy level of the signal and produces the spectrum.

Solid-state detector

A solid-state detector consists of an n-doped area of a semiconductor, in which electrons are able to move in the conduction band and a p-doped area, in which positive holes can move in the valence band. Between these areas the depletion layer is located. It is enlarged by connecting the plus pole of the voltage supply with the n-doped area and the minus pole with the p-doped area. When an ionized particle hits the detector it causes holes and electrons which move due to the electric field. These moving electrons and holes cause an electric current proportional to the energy of the particle. Because the kinetic energy of its α particle is characteristic for every decaying nucleus one is able to discriminate α decays of different parent nuclides.

Spectrum of the RAD7

After the preparations of the measurement we can pump radon containing air into the RAD7. After a short time we can see some counts in the energy interval A, which is the energy interval of the α decay of ^{218}Po. Usually the counting rate increases in the first five minutes, because in this period of time the amount of positive ionized ^{218}Po nuclei increases until it reaches a constant level on the detector. After 20 minutes the secular equilibrium state between ^{218}Po and ^{222}Rn is reached. According to chapter 2.2.3 this means, the activity of the daughter nucleus is similar to the activity of the mother nucleus. At this time almost all counts can be found in the energy level A, which you can see in figure 3.6.
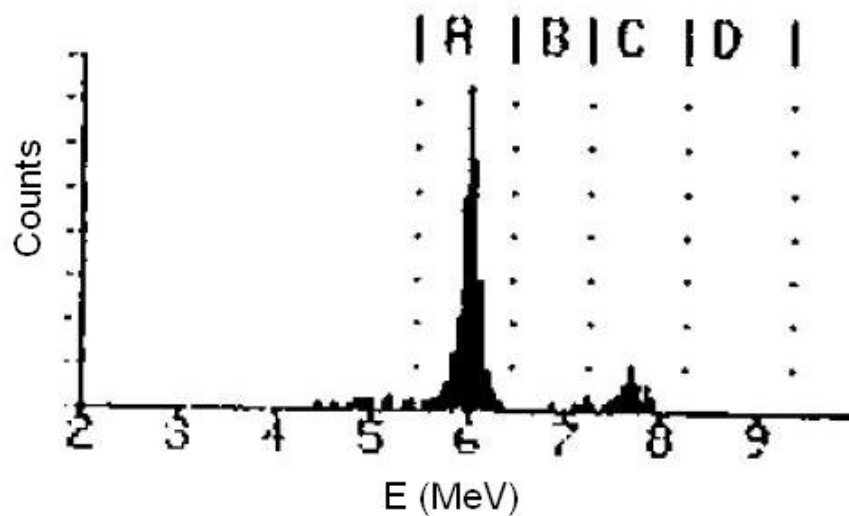
Figure 3.6: Spectrum of the RAD7 after a short period with new air which contains radon, from Durridge [2001]

After a period of time we find that the counts per time in A is constant, but the overall counting rate increase. These new counts occur at the energy level C of the spectrum. They originate from the decay of ^{214}Po (see figure 2.14) which reaches its equilibrium state after 3 hours (see chapter 2.2.3). In the full equilibrium state the height of both peaks is almost equal, as shown in figure 3.7.

Figure 3.7: Spectrum of the RAD7 after 3 hours, from Durridge [2001]; counts in A are caused by ^{218}Po and counts in C are caused by ^{214}Po

After each measurement the RAD7 has to be purged to clear the measuring chamber from the old radon concentration. In the spectrum we see that the counts with energy field A decrease fast while the counts in C stay for a longer time. The reason for this is that the counts in C originate from lead-210 and bismuth-21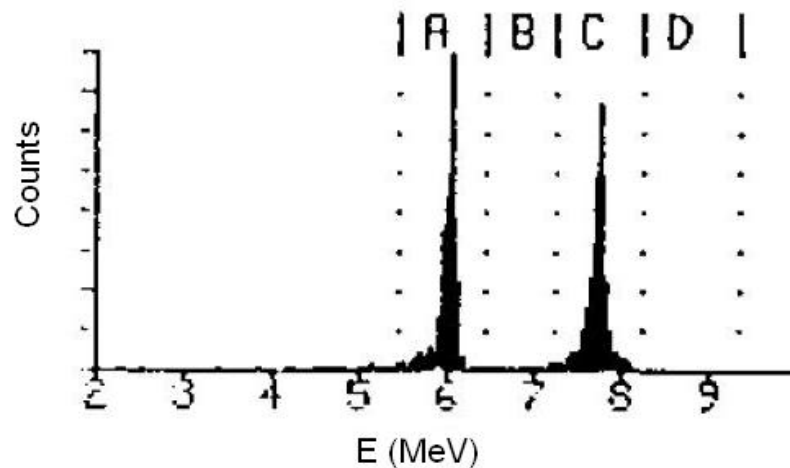4, which have much longer half lives. Usually counts in C are called „old" radon. For this practical course we will only consider the counts in A.

Notice: If the peak in C is very high before we start a measurement, we will have to take into account that the error of the RAD7 is higher because of a longer „down time" of the detector.

Efficiency of the RAD7

The efficiency of the RAD7 is very dependent on the humidity of the chamber. A bigger humidity causes a smaller counting rate. Reason for this dependency is that an ionized particle reaches a smaller range at a higher humidity because of the Bethe-Bloch-Ionization. The Bethe-Bloch-Ionization is dependent on the density of particles. If the humidity is higher, the density of particles will be bigger and the range decrease. Each time we start a measurement we have to proof that the humidity in the chamber is less than 10%. During a measurement, the sample air will be dried out with a cold trap.

3.2.2 Manual of the RAD7

This is a short version of the manual of the RAD7 from Durridge [2001]. You will find the complete version in the lab.

Main menu

The RAD7 operates with four keys which are Menu, Enter, → and ←. Have a look at the four main groups **Test**, **Data**, **Setup** and **Special** by clicking on Menu. To get into the groups

Figure 3.8: The RAD7, from Durridge [2001]

choose one with the arrow keys and confirm with `Enter`. If you want to go back just click on `Menu`.

Test includes commands to collect data series. **Data** contains commands to look at old stored data. With the help of **Setup** the parameters for the measurement can be set. **Special** will not be used during this practical course. In the following sections only the commands which are often needed are listed. For further information consult the manual of the RAD7 from Durridge [2001].

Setup

Before we can start a measurement we have to choose the right adjustment. The commands usually needed are *Setup Cycle*, *Setup Recycle* and *Setup Pump*.

We choose the time interval of one measurement with *Setup Cycle* and how often this time is repeated with *Setup Recycle*. To get a small error we usually make several measurements and take into account the error of all counts. If we decide to measure three times with a period of 15 minutes we have to set the *Cycle* to 00:15 (to get zero hours just confirm zero with `Enter`) and the *Recycle* to 3.

The different measuring techniques described in the next chapters need different adjustments of the pump of the RAD7. With *Setup Pump* we can choose between ON, OFF, GRAB or AUTO. With the help of ON/OFF the pump is permanently on/off. If we choose GRAB the pump is on for five minutes, than a five minutes equilibrium phase follows and the

RAD7 starts counting. This setup for the pump will usually be used if you want to determine the radon concentration of the air in your cellar. For the practical course the option AUTO is more relevant. If you choose AUTO the pump will start after the humidity in the chamber of the RAD7 reaches 10%. If the humidity stays less than 10% the pump will be on every five minutes for one minute until the end of the measurement.

Test

We choose the command *Test Status* to look at the current measurement. In the left upper part of the display you see the „run" and directly behind it the „cycle". For example, a displayed 0503 tells us that we measure the 5th measurement with the RAD7 and are presently in the third iteration. On the right side of the „cycle" you either see Idle or Live. Idle means the detector is in stand-by mode, Live indicates it is working. In the lower left side of the display you see the time left and in the lower right side how many counts have been detected in the current cycle.

After pressing the right arrow key once you can see the next status window. On the display you now see the data of the last finished cycle. As before you see the „run" and „cycle" in the left part. The center shows the value of the radon concentration with its statistical error (note: the RAD7 considers not only the one sigma error but further errors due to the dead time of the detector). Right behind the value you see the symbol for the unit. Usually there will be a „b" standing for Becquerels/m^3. If you see a „p" for picoCuries/litre you have to change this in the menu **Setup**.

You can get to the next status window with the arrow key. The upper left part shows the temperature. A very important information you find in the upper right part: the humidity. If the value of the humidity is above 10% please call your supervisor. In the lower left part you see the voltage of the batteries. This value should be between 6.00 V to 7.10 V. If this is not the case, call the supervisor. In the right lower part the value of the electric current of the pump is shown. The value should be between 0 and 80 mA. If this is not the case the filter is possibly blocked and needs to be changed by the supervisor.

Test Start starts the measurement and the RAD7 will automatically print and store the measurement in the end. After a measurement you have to clear the air in the RAD7 from old radon and other decay products. This is possible with the command *Test Purge*. If you want to stop the purging choose NO with the arrow key and confirm with `Enter`.

4 Questions

After reading chapters 2 and 3 you should be able to answer the following questions:

1. Which parameters do you have to consider if you want to determine the density of water?

2. How many vertical layers in a lake do you know? Do you know the names and their origin?

3. What means stratification of a lake? What is the opposite state? Can you explain the seasonal change of the temperature profile?

4. Why do we need the Brunt-Väisäla frequency? Which concept is the derivation based on?

5. What are the basic principles of the budget-gradient method?

6. What temperature, density and oxygen profile do you expect for Lake Willersinnweiher at the moment?

7. How does radon enter lake water and groundwater?

8. In which lake layers do you expect a higher radon value? Why?

9. What does secular equilibrium mean? Why is it important for this practical course?

5 Implementation

Figure 5.1: Measurements at Lake Willersinnweiher in 2008

5.1 Measurements at Lake Willersinnweiher near Ludwigshafen

5.1.1 Materials

- rubber dinghy (boat, panels, paddle)
- electric motor and a fully charged car battery
- pneumatic pump and adapter
- probe (Hydrolab) and 8x AA batteries
- laptop (fully charged battery!)
- USB cable to connect the probe to the laptop
- pipes to collect water samples (approximately 20 m depth)
- water pump (for pipes) and a fully charged car battery
- holder for pipe with a pair of pliers
- ruler
- barometer
- deionized water
- life jackets
- key to open the groundwater measuring site (GWM)
- tape and pencil
- 4x 250 ml bottles for groundwater samples
- buckets for lake water samples (4x 12 l buckets)
- key for the gate to Lake Willersinnweiher
- blue barrel to fill the buckets

For rainy days:

- umbrella
- rain wear

Note:

- in summer the easiest way to walk on the boat is barefoot
- don't forget a towel
- please take something to eat with you for the lunch break

Attention: you are only allowed to do this practical course if you can swim!!!!

5.1.2 Measurements at the lake

We will start at 9.00 o'clock on Monday morning. Meeting point is room 202 on the second floor in the Institute of Environmental Physics. Usually your supervisor will test your knowledge with questions like those in chapter 4. If you have done well you are allowed to do this practical course. You will pack the car with the help of the list of materials shown above and your supervisor will drive you to Lake Willersinnweiher (the ride takes approximately 30 minutes). At Lake Willersinnweiher you will inflate the rubber dinghy and load it.

Water level of the lake

On the lake you will measure the water level with the help of a reference level. This reference level is an old weather station. We know that the top of the steel girder is located at 88,582 m a.s.l.(= above sea level). This information is necessary for the interpretation of your data with respect to older data.
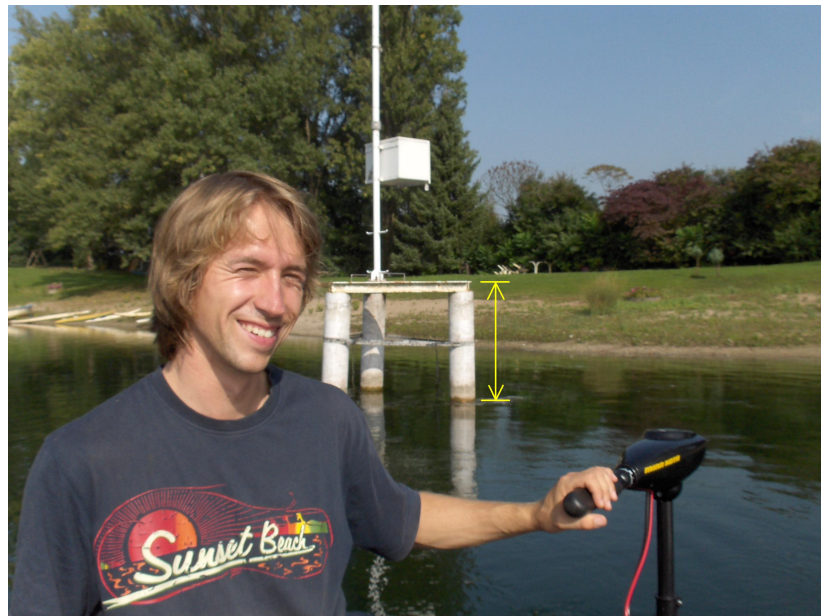

Figure 5.2: Weather station at Lake Willersinnweiher

Calibration of the probe

First step is to wheel down 20 m of the cable and insert the batteries (mind the polarity!). Then you connect the probe to the laptop with the USB cable and start the software Hydras 3 LT.

You have to calibrate the probe before you can use it. We first calibrate the oxygen sensor. Therefore, we fill deionized water to the lower part of the membrane by paying attention not to get a drop on the membrane itself. Now you put on the cap of the protecting calibration cup , so that a pressure equilibration is still possible. After a short period of time we will reach 100% humidity in the cup. Now you can measure the air pressure and convert the unit from *mbar* to *Torr*. The pressure value in the unit *Torr* has to be filled in the calibration window DO[%] and you have to click on `Kalibrierung` to calibrate. Usually you will see a

popup window with the message *Kalibrierung erfolgreich* which means the probe is calibrated. If this is not the case just try it again.

After calibrating the oxygen sensor you have to exchange the calibration cup for the measuring cage. The measuring cage will protect the sensors of the probe head in case it reaches the lake's ground. With the measuring cage you lower the probe to the water until the pressure sensor is at water level. Now you choose the menu Depth in the calibrating window and click on `Kalibrierung`. You should see a value of 0 m for the depth now. Now the probe is used as described in chapter 3.1.2.

Recording vertical profiles

After calibrating the probe and handling the software you can measure your first vertical profile. You take a profile by lowering the probe slowly and continuously to the ground of the lake with the help of the cable. After you took the first profile pull up the probe, wait a short while and let it down again in steps of 2 m, 1 m or 0,5 m. You have to wait at the different depths until the oxygen sensor has stabilised.

At the lake you have to measure five profiles: two at the deepest point of the lake (measuring site A), two at the smaller part of the lake (measuring site B) and one at the measuring site C where we take our water samples. The student who handles the laptop has to remember in which depth the thermocline is located because we will take the water samples in relation to the thermocline.
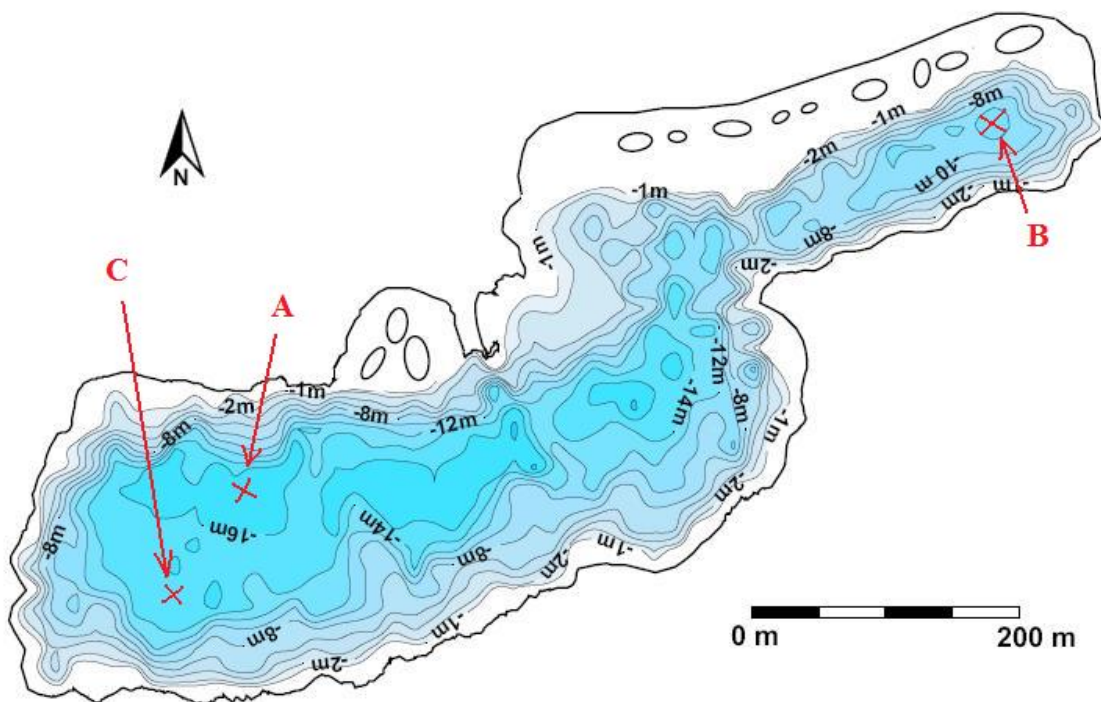

Figure 5.3: Measuring sites A, B and C at Lake Willersinnweiher, from Wollschläger [2003]

Taking lake water samples

Take lake water samples from at least three different depths using the pipes and the water pump. You have to pump a while until you can be sure that the water you sample is from the corresponding depth. When filling the buckets take care that there are no air bubbles inside. This is very important because if we have air bubbles in the sample we will loose radon according to Henry's law.

Taking groundwater samples

After taking lake water samples you have to pack everything into the car again and we drive to the groundwater measuring site. There you will collect 2 x 250 ml samples at two different depths. It is important to have no air bubbles inside your samples, as mentioned before. With the help of the probe you determine the level of the groundwater and take a vertical profile. We know that the top of the groundwater pipe is located at 92,3 m a.s.l.

Figure 5.4: Groundwater measuring site B at Lake Willersinnweiher

5.2 Measurements in the hydrology lab

The practical course F50/51 always takes place with four students. One group will start with the lake water measurement just after the field trip on Monday afternoon. The other group will start the lake water measurement Tuesday afternoon. We start with the lake water samples because the radon activity concentration is much lower than in groundwater. On Wednesday both groups will do the groundwater measurement. Because the duration of the measurements of lake water samples is much longer than that of the groundwater analyses, the former are run overnight. Due to didactical reasons the groundwater measurement is introduced first and afterwards the lake water analysis is adressed. You can do the interpretation of the profiles during the radon measurements.

Materials

- RAD7
- water samples from the lake and groundwater
- water pump
- RAD H_2O
- cold traps
- some plastic pipes and adapters
- dry ice-isopropanol mixture
- exchanger
- thermometer
- scales

5.2.1 Groundwater measurement with the method RAD H_2O

Setup RAD H_2O

To determine the radon concentration in the groundwater samples the RAD H_2O is used (see figure 5.5). This setup is able to measure radon activity concentrations in the range above $100 \, \mathrm{Bq/m^3}$. The RAD7 is not able to determine the radon concentration of our water samples directly because it can only handle gases. Thus, an equilibrium between a certain gas volume - we use air - and a known water volume has to be established first. For this purpose we use a glass frit (shown in figure 5.6). According to the manufacturer Durridge about 95% of the radon passes to the gas phase within 5 minutes. The final concentration of radon in the gas and the fluid phase can be described by Henry's law.

After reaching the equilibrium state the air is dried with a cold trap and is then pumped into a closed loop to the RAD7. Knowing the volume of the water and the volume of the air in the closed loop we can calculate the activity concentration in the water sample from the measured activity concentration in air.

Calculation of the radon activity concentration

To reconstruct the radon activity concentration in water, $C_{W,0}$, from the measured concentration in air we have to know the volumes of the water and the air loop (V_W and V_{Air}) and the water temperature, T, at the time of the measurement. At the beginning all radon is located in the water ($C_{W,0}$). After reaching the equilibrium state the activity concentration is allocated in both phases (C_{Air} and C_W):
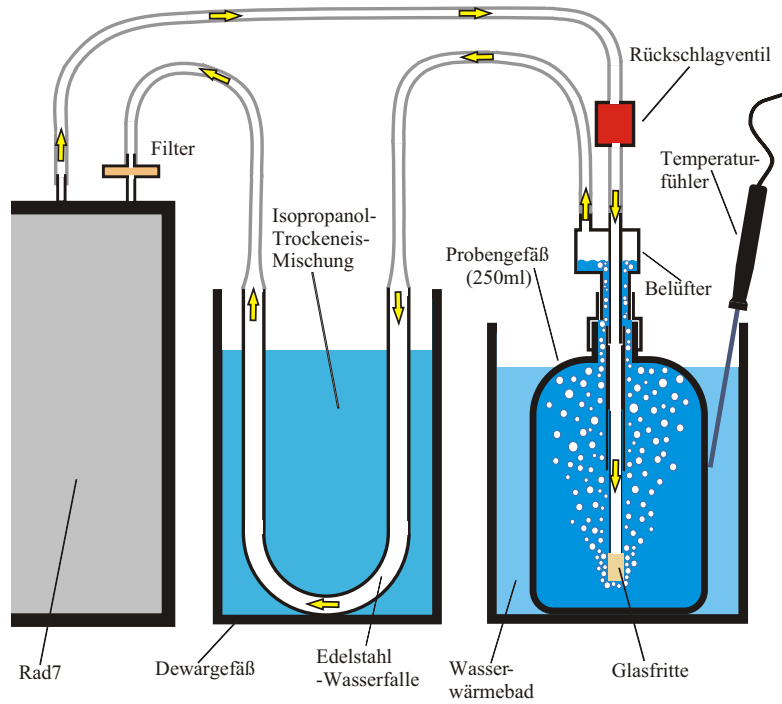
Figure 5.5: Overview of RAD H$_2$O, from Reichel [2009]

$$C_{W,0} \cdot V_W = C_{Air} \cdot V_{Air} + C_W \cdot V_W \qquad (5.1)$$

We have to take into account that the RAD7 always has a small underground radon concentration C_U:

$$C_{W,0} \cdot V_W + C_U \cdot V_{Air} = C_{Air} \cdot V_{Air} + C_W \cdot V_W \qquad (5.2)$$

The activity concentration in water in the equilibrium state can be described by equation 2.50. As mentioned in chapter 2.4.3 the Ostwald solubility coefficient is dependent on temperature:

$$C_W = a(T) \cdot C_{Air} \qquad (5.3)$$

With the help of equations 5.2 and 5.3 we can calculate the activity concentration in water at the time of the measurement. This value has to be corrected after the radioactive decay law, because a part of the radon in the water sample has already decayed during storage time.

Measuring groundwater

For the method RAD H$_2$O we start with an underground measurement. For this purpose you connect everything as described above, but take an empty bottle instead of a full one

Figure 5.6: Glass frit, from Durridge [2001]

(30 minutes measuring time). After the underground measurement change the bottles, choose a 20 minutes cycle and switch the pump to *On* to reach the equilibrium state. After equilibrium is reached choose a measuring time of 60 minutes (pump on *Auto*). During the measurement don't forget to note the temperature.

As mentioned before you have to make sure that during the whole measurement no water gets into the RAD7! After the measurements you have to purge the cold traps and the RAD7.

5.2.2 Lake water measurement with the method RAD Aqua Plus

The radon activity concentration of lake water (range of 5 to 30 Bq/m^3) is much lower than the one of groundwater (range of 5 000 to 10 000 Bq/m^3). The errors of RAD H$_2$O method are too large for the small activity values of lake water. Therefore, people at the Institute of Environmental Physics have developed new methods to determine the radon activity concentration of lake water with an acceptable error in short time.

Setup RAD Aqua Plus

The method RAD Aqua Plus - as the method RAD H$_2$O - is based on the equilibrium between a closed water loop and a closed air cycle. Figure 5.7 shows the method in principle.

The lake water samples are taken in 12 l buckets. Before the measurement the normal top is exchanged by a prepared top with an integrated pump. The water pump is controlled by a power supply. The pumped water is sprayed into the exchanger in which the contact between water and air loop takes place. Afterwards, the radon containing air is pumped through the cold trap to dry out and then into the RAD7. The equilibrium state is reached after 40 minutes at the given pump rate for the RAD Aqua Plus setup.

Take care that the exchanger is completely empty. Any residual water distorts the measurement. Furthermore check that the one-way valve points in the right direction.
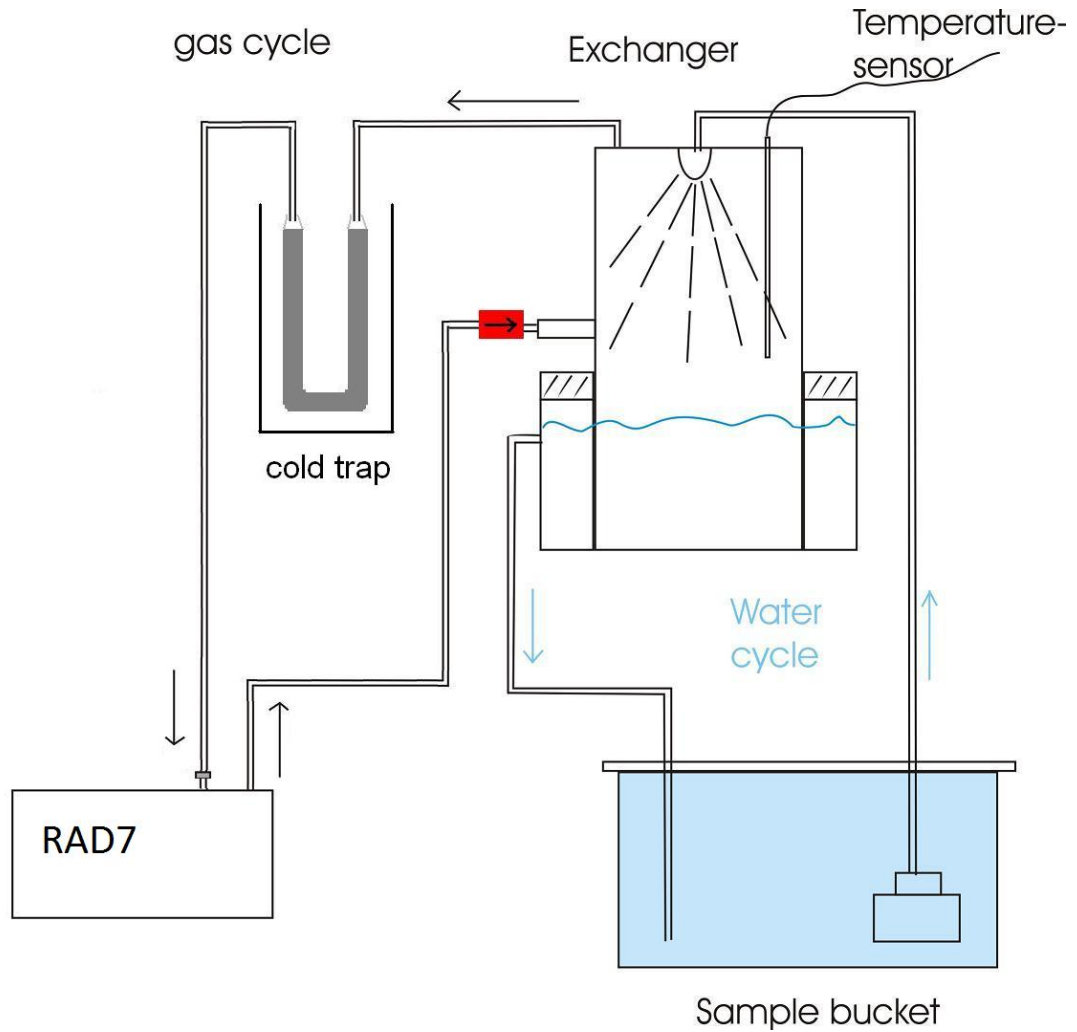
Figure 5.7: Structure of RAD Aqua Plus, from Kluge et al. [2007]

Calculation of the radon activity concentration

The calculations are based on the same ideas as the calculations of method RAD H_2O. Therefore, we get the following equation:

$$C_W = C_{Air} \left(\frac{V_{Air}}{V_W} + a(T) \right) \tag{5.4}$$

with

C_W	activity concentration of the water phase
C_{Air}	measured activity concentration in air
V_{Air}	measured volume of air in the closed gas cycle
$a(T)$	Ostwald solubility calculated according to equation 2.51

The last step is, again, to correct the value C_W with the help of the radioactive decay law.

Measuring lake water

First step: connect everything and open or close the valves in the air loop (see figure 5.8).
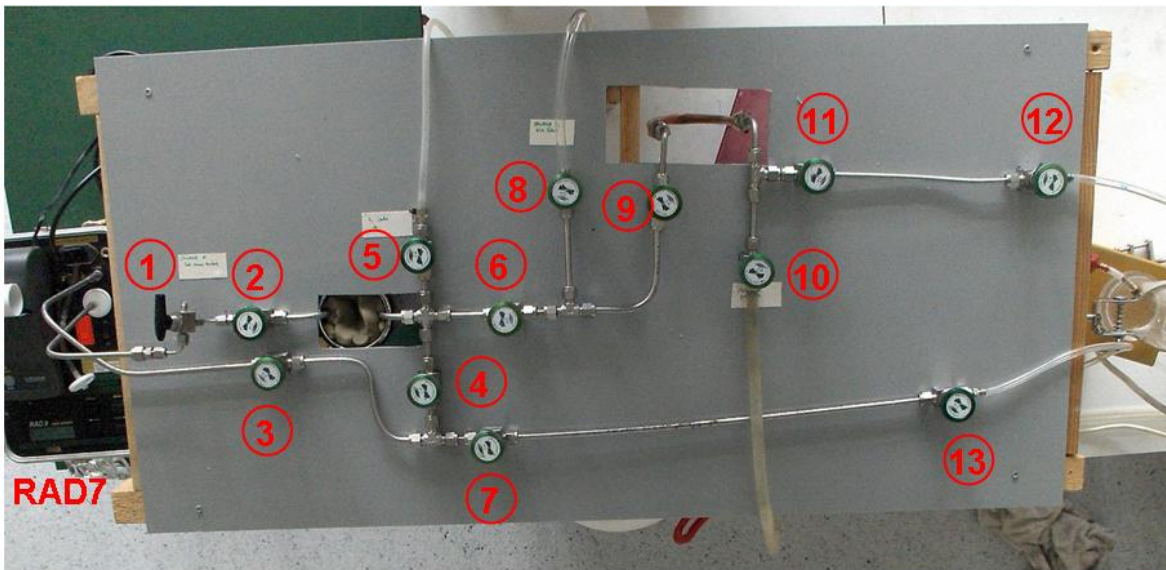

Figure 5.8: Rad Aqua Plus in the lab

Before we start the measurement we have to purge the whole construction with nitrogen for 10 minutes to assure a low background signal. For the purging with nitrogen the valves 8, 4 and 10 have to be closed and the others have to be open. Switch 1 has to be switched in that way that nitrogen reaches the RAD7.

While the air circle is being purged with nitrogen you can determine the weight of the buckets and the water temperature. After you have finished purging the air loop you have to make sure that switch 1 lets the air to the RAD7, that the valves 2, 3, 7, 13, 6, 9, 11 and 12 are open while the others are closed. After checking you exchange the normal top with the measuring top of the bucket and regulate the water flow of the pump. Now you have to set up the RAD7 to five cycles with 10 minutes each and the pump to *On* immediately.

During the measurement the water level in the exchanger should be at the same level as the tape. You have to look after the water level very carefully because the exchanger must not fill up with water completely. This would enable the water to reach the RAD7 and damage it heavily.

After reaching the equilibrium state we will measure the radon in the air only in a small air cycle without the exchanger. Therefore, you should turn off the pump of the RAD7, close the valves 6 and 7 and open valve 4. Afterwards you put the pump on *Auto* and set up three additional cycles with at least four hours. Now you measure the water temperature again and empty the exchanger for the next group. After all measurements are finished you have to clean the cold traps from ice.

Some data for the analysis

	Volume	Weight
Bucket	12 l	0.35 kg
Bottle	250.0 ± 5.0 ml	0.17 kg

Table 5.1: The sampling containers

	RAD7	Volume
RAD H_2O F50	1317	899 ± 3 cm^3
	2224 / 2225 / 2409	1092 ± 3 cm^3
RAD H_2O F51	1317	897 ± 3 cm^3
	2224 / 2225 / 2409	1094 ± 3 cm^3
RAD Aqua Plus	1317	1608 ± 200 cm^3
	2224 / 2225 / 2409	1800 ± 200 cm^3

Table 5.2: Volumes of the experimental set-up

RAD7	Conversion factor
1317	151,7 Bq/m^3 = 1 cpm
2224	59,7 Bq/m^3 = 1 cpm
2225	61,4 Bq/m^3 = 1 cpm
2409	62,5 Bq/m^3 = 1 cpm

Table 5.3: Conversion factors for each particular RAD7

6 Working tasks

In sections 6.1 and 6.2 you will find tasks which you have to deal with. Please note that all students have to deal with tasks (1.1), (1.2) and (1.3) for the limnological interpretation of the data and can do tasks (1.4) and (1.5) additionally. Further note that tasks (2.1) and (2.2) for the groundwater lake interaction have to be done by all students and (2.3) and (2.4) can be dealt with additionally.

6.1 Limnological parameters

Tasks:

(1.1) Draw and interpret the profiles of temperature, conductivity oxygen and density in dependency of depth.
Where are the different layers? What are the differences between A and B?

(1.2) Compare your profiles with older data and show the seasonal change (For example, put some older profiles with yours into one plot).

(1.3) Calculate the vertical stability and interpret your results.

(1.4) Calculate the mixing coefficient with the help of the budget-gradient method for different depths and interpret your results.

(1.5) What is the dependency of stability and mixing coefficient? (Determine with the help of a figure)

6.2 Groundwater-lake interaction

Tasks:

(2.1) Calculate the different radon activity concentrations for the different methods.

(2.2) Plot your measurements and interpret your results.
In which depth is groundwater located?
Does this information give evidence of the depth of infiltration into the lake?

(2.3) Compare your radon profile with older ones.
What is the reason for the differences?

(2.4) Measuring techniques: why do we have to use different methods for groundwater and lake water? What are the advantages/disadvantages of each method?

Important notes:

Please note your results in the lab book that succeeding groups can refer to your data. Make sure that you work with a copy of the original data!

7 Notes to the tasks

In the lab you find a lab book. Please write down your results into the lab book so that succeeding groups can refer to the data. In the following sections you get hints for the working tasks. Please use these hints only, if you have no own ideas for the solution.

7.1 Analysis of the limnological parameters

Please copy the original data into a file with your names (folder: Eigene Dateien/FP Limno-physik/your names). In this file you can work with the data and store your figures and origin projects.

7.1.1 Notes to task (1.1)

In the following sections you will see how to work with the data of the probe. First step is to clear the table of a profile from the entries you do not need. This means you should delete the lines in the table in which you see that the probe was not in the water yet. Further, you should look at the end of the table and delete the lines in which the probe lay at the ground. The last column contains the circulator status. Make sure the status of the circulator is 1 in every depth. If this is not the case keep the column and include it in your interpretation, otherwise you can delete the column.

From the corrected table we only need the columns of depth, conductivity and temperature. You can delete the other columns. In this table you can calculate the values of the reference conductivity with the help of equation 2.10. The density of lake water is calculated in two steps. Step one is to calculate the density of pure water with equation 2.2 and step two is to calculate the density of lake water with the density of pure water and the reference conductivity with equation 2.13.

Open this table with the calculated values with the software Origin to plot the profiles. Please note that we are drawing profiles which means that the depth z is always located on the y axis (be careful with the direction).

After these profiles are drawn you can open the second data set with the oxygen data. We want to determine the oxygen value in mg/l for the different depths. Therefore, we plot the oxygen value of one depth in dependency of time. How do you get the oxygen values? Please make a short note how you determined it. At the end you should have gotten the oxygen values for the different depths and plot an oxygen profile. Is it necessary to plot the error bars?

Perform these evaluations for both measuring sites A and B. Where are the differences between the two profiles? What is similar? How can you explain the shapes of the profiles?

If you want to print the profiles you have to export the pictures from Origin by choosing *Datei* and then *Grafik exportieren.*

7.1.2 Notes to task (1.2)

We want to compare profiles from different dates. Therefore, we have to convert the depth in m to the depth in m a.s.l. For this conversion you need the distance from the water level to the steel girder (88,582 m a.s.l.). Please copy your values into the data collection you find on the PC in the file FP Limnophysik. Pay attention to copy the values of the table and not the functions.

You can plot more than one profile in one graph with Origin. Therefore, you have to open the complete data table and choose *Liniendiagramm* in the menu *Zeichnen*. Afterwards you choose the x and y axis of one date and click on *hinzufügen*. You have to continue with the next dates in the same way. After plotting all the available data sets choose the ones you really need and print them. Did you expect this seasonal change? Give reasons!

7.1.3 Notes to task (1.3)

Calculate the stability frequency as a differential quotient with the help of equation 2.22. After calculating the stability values you can average over five values to decrease the variabilities. You have to consider that there are positive and negative values for the stability frequency if you plot a graph in a logarithmic scale. You should decide whether you want the stability N^2 in linear or logarithmic scale and add the density to the graph. Afterwards, it is easy to interpret the graph.

7.2 Analysis of the groundwater-lake interaction

7.2.1 Notes to task (2.1)

Calculation of the radon activity concentration of lake water

The first step is to calculate the Ostwald solubility with the help of the measured temperature. Do not forget to consider the error in the temperature! Therefore, look at equation 2.51 and use the propagation of uncertainty from Gauss.

The second step is to get the volume of the water. Therefore, you have to weigh the buckets with the water and subtract the mass of the empty buckets. Afterwards, you can use the approximation that 1 kg water is equal to 1 l. Why can we use this approximation? Next step is to look up the volume of the air from table 5.2.

Now we determine the activity of the gas cycle. Therefore, you have to add the counts from energy interval A (to get the counts in A multiply the complete counts with the per cent value) and calculate the statistical error. This value (with error!) has to be corrected for the background. You find a background measurement for each RAD7 we usually use. How do you determine the background you need for your individual measurement time? Write down how you get your background and the according error. Afterwards, you have the radon activity concentration of the gas cycle in the unit counts per minute (cpm) which you have to transfer into Bq/m^3 (see table 5.3).

Next step is to calculate the activity concentration in the water phase at the time of the measurement with equation 5.4. Use the Gaussian error propagation to get your error. Which errors have to be taken into account? Which are negligible?

The last step is to take into account that time has passed since we took the samples. Therefore, you have to correct the activity concentration with the help of the decay law.

Calculation of the radon activity concentration of groundwater

The first step in this part is similar to the first step in the last section: we have to calculate the Ostwald solubility $a \pm \Delta a$. The next step is much easier, because in the range of groundwater the RAD7 calculates the right value of the radon activity concentration in the gas phase. You just have to consider the background measurement and calculate the error. Usually the duration of your background measurement and the real measurement is not equal. Do you have to consider this difference in time?

The last two steps are equal to the lake water calculation: determine the radon activity concentration in the water phase and consider the time difference using the radioactive decay law. After calculating the values of groundwater and lake water you are able to answer the following question: is radon a good tracer?

7.2.2 Notes to task (2.2)

In the lab you will find a lab book. Please write down your radon calculations and the other information which is necessary to interpret the values. In the lab book you can look up the last radon measurements. Plot these data with your radon data to get a better radon profile. What can you see in the profile? How can you interpret your graph? Are there any relations between the groundwater level and the radon activity concentrations? Reasons? Figure 7.1 shows a radon profile of Lake Willersinnweiher in summer 2005. Does this profile match your results? What are the differences?
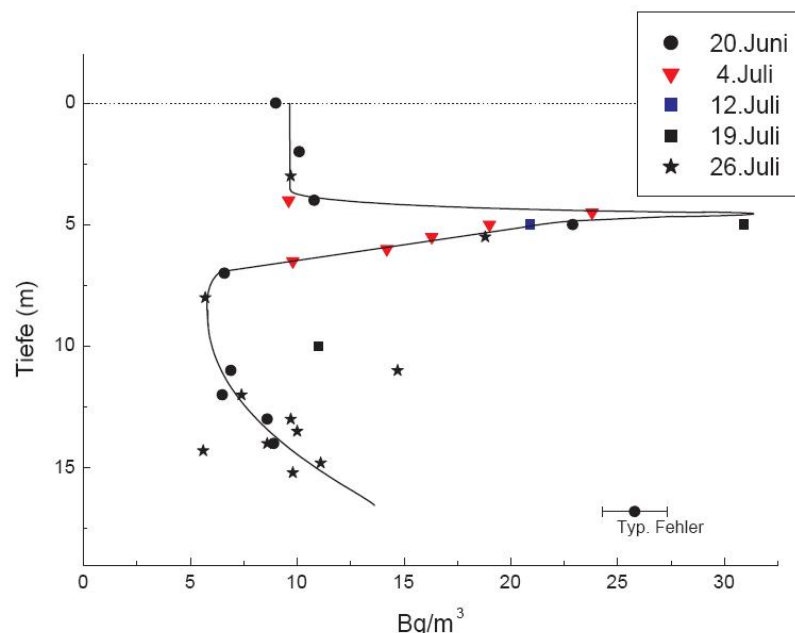

Figure 7.1: Radon depth profile of the Lake Willersinnweihers in summer 2005, from Kluge [2005]

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