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Sensitivity of Ancient Lake Ohrid to Local Anthropogenic Impacts and Global Warming

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ABSTRACT. *Human impacts on the few ancient lakes of the world must be assessed, as any change can lead to an irreversible loss of endemic communities. In such an assessment, the sensitivity of Lake Ohrid (Macedonia/Albania; surface area A = 358 km^2, volume V = 55 km^3, > 200 endemic species) to three major human impacts—water abstraction, eutrophication, and global warming—is evaluated.*

It is shown that ongoing eutrophication presents the major threat to this unique lake system, even under the conservative assumption of an increase in phosphorus (P) concentration from the current 4.5 to a potential future 9 mg P m^{-3}. Eutrophication would lead to a significant reduction in light penetration, which is a prerequisite for endemic, deep living plankton communities. Moreover, a P increase to 9 mg P m^{-3} would create deep water anoxia through elevated oxygen consumption and increase in the water column stability due to more mineralization of organic material. Such anoxic conditions would severely threaten the endemic bottom fauna. The trend toward anoxia is further amplified by the predicted global warming of 0.04°C yr^{-1}, which significantly reduces the frequency of complete seasonal deep convective mixing compared to the current warming of 0.006°C yr^{-1}. This reduction in deep water exchange is triggered by the warming process rather than by overall higher temperatures in the lake. In contrast, deep convective mixing would be even more frequent than today under a higher temperature equilibrium, as a result of the temperature dependence of the thermal expansivity of water. Although water abstraction may change local habitats, e.g., karst spring areas, its effects on overall lake properties was shown to be of minor importance.

INDEX WORDS: *Eutrophication, water abstraction, global warming, Lake Ohrid, aquatic ecosystem, ancient lake, endemism.*

INTRODUCTION

The following paper assesses the effects of three major human impacts—eutrophication, water abstraction, global warming—on the physical properties of Lake Ohrid (Table 1). Particular emphasis is given to the possible interaction between the local and global human impacts, as well as their relative importance. The analysis focuses on properties that form physical habitats for the endemic species of the lake.

Lake Ohrid, located between Macedonia and Albania (Table 1, Fig. 1), is among the few ancient, long-lived lakes of the world that have provided continuous freshwater habitats for more than 1 million years, and the only one in Europe (Meybeck 1995, Gorthner 1994, Martens *et al.* 1994). It harbors a large number of relict and endemic species, which makes it also a hotspot of freshwater diversity (UN: World Conservation Monitoring Centre 1998, LakeNet: Duker and Borre 2001). Its importance was further underlined by UNESCO, by declaring it as a World Heritage site (UNESCO 1979).

While Lake Ohrid is currently oligotrophic and not in immediate danger, it is being jeopardized by human activities (Watzin *et al.* 2002). In the two riparian countries, which are both in political transition, agriculture is still a major source of economy, putting pressure on surface water for irrigation given the semi-arid climate of the area (European Environment Agency 2003). The lack of economic

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TABLE 1. Characteristics of Lake Ohrid.

Property	Unit	Value
Latitude[1]	° N	41.1
Longitude[1]	° E	20.8
Altitude[1]	m asl	693.7
Catchment area[2]	km^2	2600
Surface area[3]	km^2	358
Volume[3]	km^3	54.9
Maximal depth[3]	m	288.7
Average depth[3]	m	155
Hydraulic water residence time[3, 4]	yr	70
Age of lake existence[5]	10^6 yr	2–3
Endemic species[5]	#	> 200
Average phosphorus concentration[6]	mg-P m^{-3}	4.5

[1]Naumoski (2001)
[2]Including Lake Ohrid tributaries (Watzin *et al.* 2002), as well as Lake Prespa and its catchment (Anovski *et al.* 2001)
[3]Based on bathymetric data measured by the Smithsonian Institution (unpublished data, 1975)
[4]Macedonian Hydrometeorological Institute (unpublished data)
[5]Stankovic (1960)
[6]Own measurements (2002–2004).

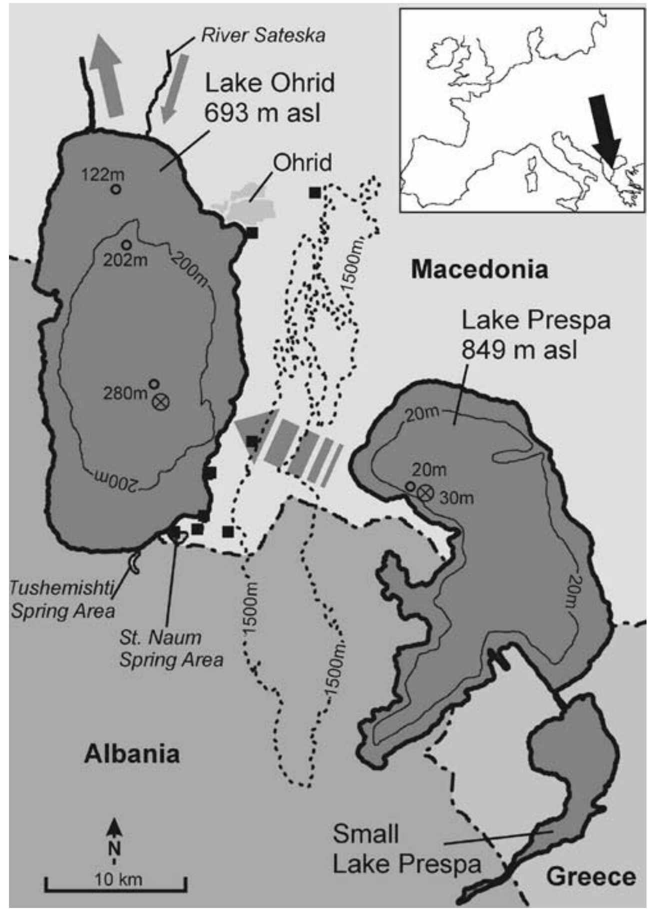

FIG. 1. Geographical overview. Gray arrows show water flow (broken arrow = underground flow), crossed circle is main sampling site, inset map indicates location of study area in Europe.

growth also leads to pollution from sewage, because of missing or ineffective wastewater treatment (Ernst Basler and Partners 1995). The local human impacts are aggravated by an ongoing growth in population within the lake catchment (Albania: Watzin *et al.* 2002; Macedonia: Macedonian State Institute for Statistics, personal communication 2004). Moreover, regional simulations of global warming predict changes above average for south-eastern Europe (Räisänen *et al.* 2004, Giorgi *et al.* 2004). Finally the large depth of Lake Ohrid (Table 1) makes it prone to prolonged deep water isolation due to temperature increase (IPCC 2002, O'Reilly *et al.* 2003).

The following case study provides an example of a system-analytical assessment, applicable to many other lakes that are exposed to similar threats. Indeed, pollution and scarcity of freshwater resources, followed by climate change, have been identified as the major environmental concerns of the future (UNEP 1999). Several recent publications have addressed the effect of these three environmental threats—pollution, water abstraction, and climate change—on physical lake processes (e.g., Goudsmit *et al.* 2002, Sirjacobs *et al.* 2004, Lehmann 2002). Anthropogenically-induced changes in the physical lake properties have in turn been described to affect lake ecology (O'Reilly *et al.* 2003, Hecky *et al.* 1994) as well as economy (Lofgren *et al.* 2002). While some of the observed or predicted changes in lakes have been shown to be significant (Beeton 2002), comparison of impacts and interactions among them are often neglected. Impact assessment of these human pressures deserves special attention for the few globally unique ancient lakes, as any change might lead to irreversible losses (Beeton 2002, Cohen 1994).

Summarizing the above, there are three main motivations for our work:

(a) Lake Ohrid is an ideal site to study the influence and interaction of the three human impacts, water abstraction, eutrophication, and climate change on physical lake properties.

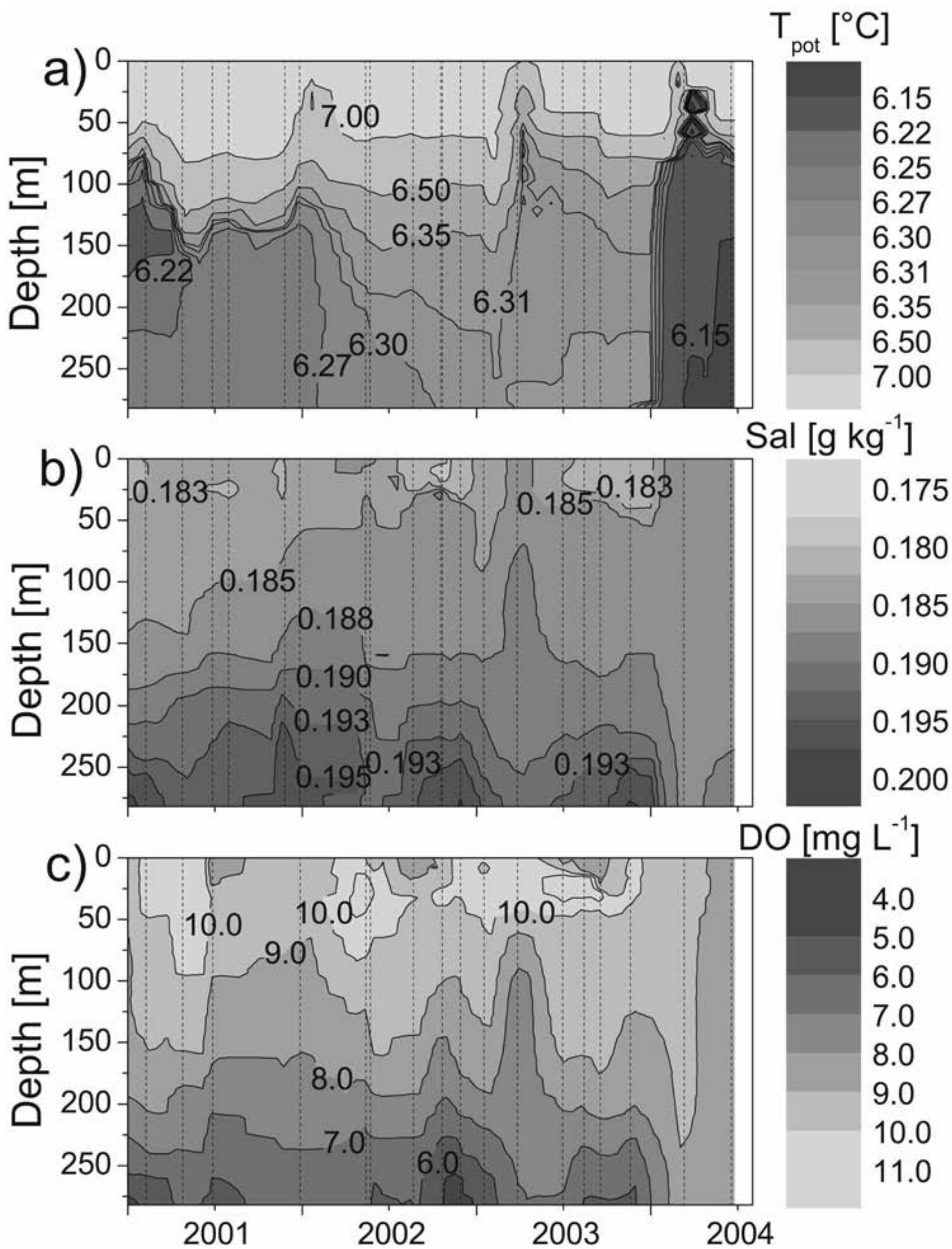

FIG. 2. Contours from vertical profiles over the period 2001-2004. Dotted lines indicate dates of vertical profiles. (a) Temperature measured with a CTD probe, corrected for the surface level. Note that surface temperatures reach up to 27°C in summer, which is not resolved with the chosen scale. (b) CTD conductivity transformed into salinity based on the average salt composition of Lake Ohrid. (c) Dissolved oxygen (DO) concentration from CTD measurements. Each DO profile is calibrated with a profile of Winkler samples.

(b) Its endemic communities are threatened by potential changes in the physical boundary conditions.
(c) In view of its international importance, the lake system is poorly documented, especially in terms of physical processes. A description of its status quo is required as a reference for future evaluations of changes.

APPROACH

The main management goal for Lake Ohrid is to conserve its rich and extensive endemism. We here base our analysis on properties, which form the physical habitat for two important groups of endemic species, phytoplankton and bottom fauna.

Phytoplankton

During the main productive season from April to October vertical phytoplankton distribution in Lake Ohrid is quite typical for oligotrophic lakes with green algae (*Chlorophyta*, *Chrysophyta*, *Pyrrophyta*) dominating the top 10 m of the water column and small forms of *Cyanophyta* taking over between 10 and 30 m (Patceva 2001, Kalff 2002). However the lake harbors highly specialized forms of pelagic diatoms (e.g., *Cyclotella fottii*) which show major growth between 20 and 50 m depth and are clearly dominant between 40 and 150 m (Patceva 2005, Mitic 1985, Stankovic 1960). In the oligotrophic lake (Table 1) high competition for bio-available soluble reactive phosphorus (SRP) is expected (N:P > 25:1, Matzinger *et al.* 2005). Thanks to the efficient light use and/or the large size (up to 1 mm) of endemic phytoplankton species (Mitic 1985), they are able to populate deep layers where most nutrients are available. This is also indicated by the shift of their maximal density to a depth of ~75 m from late autumn to early spring when the water has its highest clarity (Patceva 2005). Two criteria emerge: first the endemic diatoms depend on good light conditions at large depths and thus on high water transparency; and second, they are dependent on the supply of SRP between 20 and 150 m depth.

Bottom Fauna

About 90% of the endemic fauna described in Lake Ohrid are benthic organisms (Stankovic 1960). While some are found throughout the lake, most of them are limited to a specific depth zone. A surprising number of species is even restricted to large depth exceeding 150 m, such as some forms of gastropods, amphipods and ostracods (Stankovic 1960). Other species, such as the peculiar sponges *Ochridaspongia rotunda* and *Ochridaspongia interlithonis,* live in deep water or at shallow sites in the vicinity of subaquatic spring inflows (Arndt 1937, Gilbert and Hadzisce 1984). Both areas are defined by cool and oxygen-rich water. Given the oligotrophic conditions all the endemic forms favor oxygen-rich conditions and are expected to be vulnerable to sediment anoxia.

As a result we assessed the effects of the three major human impacts—eutrophication, water abstraction, global warming—on (i) SRP supply between 20 and 150 m depth, (ii) water clarity in the trophogenic layer, and (iii) oxygen supply to the hypolimnion.

The paper starts with a presentation of the status quo of physical properties of Lake Ohrid, which are important to understand the changes from human impacts. This description is based on basic parameters from regular CTD casts (Fig. 2, Seabird SBE 19), thermistor measurements (Vemco and RBR Ltd.) and water samples for dissolved oxygen (DO) from 1999–2004. Following the physical lake characterization, human impacts are defined for historic, current, and potential future situations. Finally, based on the physical properties, the effect of the defined anthropogenic changes is assessed.

PHYSICAL CHARACTERISTICS OF LAKE OHRID—STATUS QUO

Hydrology

Water Balance

The relatively dry, Mediterranean climate and the small drainage basin (catchment/lake surface ratio of ~7) results in a long hydraulic residence time scale of ~70 yr (Table 1). The water balance is dominated by the inflow from karst aquifers (~50%) with smaller shares from rivers and direct precipitation (Table 2). The river runoff was even lower by 5.5 m^3 s^{-1} (~70%), before 1962 when River Sateska was deliberately diverted into the lake from the north (Fig. 1). The karst aquifers are charged from mountain range precipitation and from Lake Prespa, as revealed by using stable isotopes and Cl^- as natural tracers (Fig. 1; Anovski *et al.* 1992, Eftimi and Zoto 1997, Matzinger *et al.* 2006) and dye tracer experiments (J. Zoto, personal communication 2003). Apart from two large spring

TABLE 2. Water balance of Lake Ohrid.

Outputs	Flow rate [m^3 s^{-1}]
Surface outflow River Crn Drim[1,2]	24.9
Evaporation (1145 mm yr^{-1})[3]	13.0
Total output	37.9
Inputs	**Flow rate [m^3 s^{-1}]**
Precipitation on lake surface (773 mm yr^{-1})[3]	8.8
River inflows:	
Albanian catchment[3]	0.5
Macedonian catchment without R. Sateska[4,5]	1.9
R. Sateska (diversion into Lake Ohrid in 1962)[1,2,4]	5.5
Temporary inflows (estimation)[6]	1.0
Surface spring inflows[3]	10.3
Sublacustrine springs (from closing the balance)	9.9
Total input	37.9

[1] Ivanova (1974)
[2] Macedonian Hydrometeorological Institute (unpublished data, 2001)
[3] Watzin *et al.* (2002)
[4] Hydrobiological Institute Ohrid (unpublished data)
[5] Own measurements (2002–2003)
[6] Several small streams only develop after rain storms.

areas, which flow directly into Lake Ohrid on its southern shore (Fig. 1), about 50% of the aquifers are expected to be sublacustrine (Table 2). It is noteworthy that 50% of the already small catchment area is made up of the underground connection to Lake Prespa (Fig. 1). The water leaves Lake Ohrid by surface outflow in the north (~60%) and by evaporation (~40%). Annual evaporation from the lake surface exceeds direct precipitation (Table 2).

Riverine and Spring Water Intrusions

Most of the tributaries—given their relatively small discharge of usually less than 1 m^3 s^{-1}—are entrained directly into the surface water near the river inlets. During summer seasons even River Sateska, which contributes up to 13 m^3 s^{-1} after continuous rain in winter and spring, is no more than a trickling creek, because of the dry climate and extensive upstream irrigation (Fig. 3a). However during the cold season River Sateska can plunge deep due to its higher salinity and lower temperature (Fig. 3b).

The inflow from the numerous karst springs

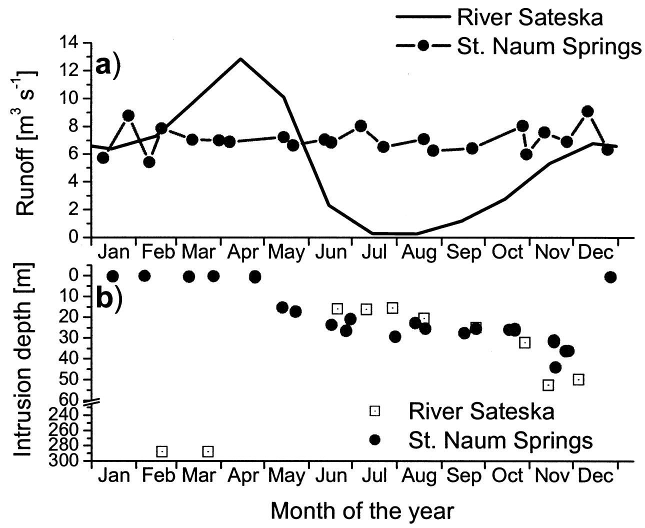

FIG. 3. Seasonal discharge dynamics of two major inflows to Lake Ohrid. (a) Runoff based on combined measurements from 1996, 1997, 2002 and 2003 for St. Naum Springs (Hydrobiological Institute, unpublished data) and data from 1949–1969 (Ivanova 1974) and 1996–2000 (Hydrometeorological Institute, unpublished data) in combination with single measurements directly at lake inflow (Hydrobiological Institute, unpublished data) for River Sateska. (b) Calculated intrusion depth based on CTD profiles for the lake and salinity and temperature measurements from the inflows.

seems very constant throughout the seasons. The surface spring area of St. Naum (Fig. 1) shows standard deviations of the flow of only ~10% (Fig. 3a), indicating significant storage capacity within the karst mountain. Though water property varies among different springs (e.g., temperature range ~9 to 12°C), each single spring area shows remarkably constant parameters. Monthly measurements in springs, one in the north-east (Biljani Spring) and one to the south-east (St. Naum Spring) (Fig. 1), showed a standard deviation of ~2% in temperature (T), ~1% in pH and ~10% in salinity (S) (Hydrobiological Institute, unpublished results, 2000 to 2002 for pH and T, 2002 for S). Based on the stable water properties, the intrusion pattern of the spring water is also very constant within a season. In general the salinity of the groundwater is higher than that of the the lake by ~0.06 g kg^{-1}. Large surface springs, entering the lake with higher salinity and lower temperature than ambient lake water will plunge close to the level of neutral buoyancy into

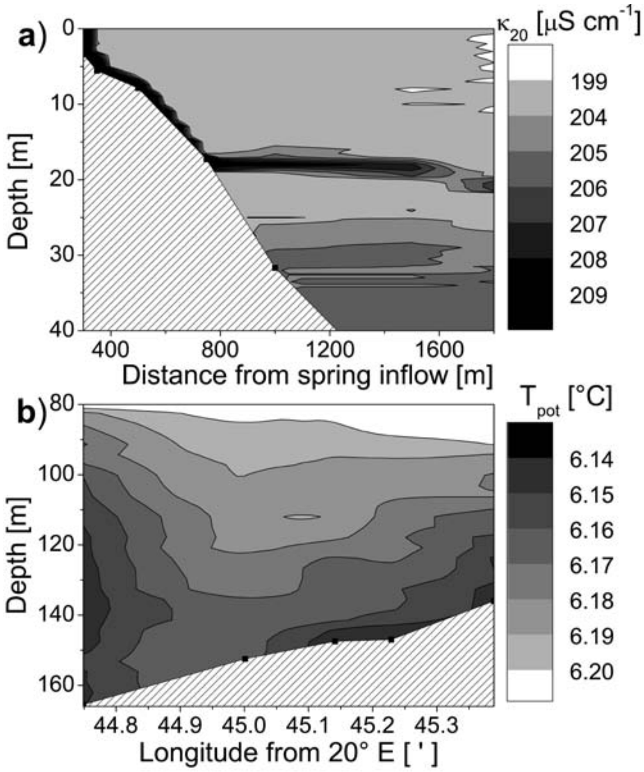

FIG. 4. Mapping of spring inflow based on CTD transects. (a) Inflow from St. Naum Springs (location in Fig. 1) plunging from the surface in terms of conductivity κ_{20} on 18 Oct. 2002. (b) Temperature transect over a subaqueous spring area in the Northern part of Lake Ohrid on 22 Jun. 2004. Hatched area is lake bottom.

the thermocline. For the large spring area of St. Naum with an average flow of 7.5 m^3 s^{-1} (Watzin *et al.* 2002, Table 2), the intrusion depth varies during the summer season from about 15 to 40 m (Figure 3b). Despite entrainment of ambient water a pronounced conductivity is seen kilometers from the spring inflow after several months of almost constant stratification in autumn (Figure 4a).

In winter the spring water is warmer and thus lighter than ambient (6 to 8°C) and merges with the surface layer (Fig. 3b). The situation is similar for the deep subaquatic springs, which enter the lake in a zone with temperatures between 6 and 7°C throughout the year. As a consequence the warmer spring waters will rise and form plume-like structures. One such plume with distinct rise of the isotherms by up to 50 m over a short horizontal distance of ~200 m was observed in 150 m depth (Fig. 4b). This same plume was detected on several instances over one year indicating continuous supply. A model approach by McGinnis *et al.* (2004) further verifies that subaquatic spring inflows would indeed create a structure similar to the one observed (Fig. 4b).

Stratification and Mixing

Stratification

Whereas the surface temperature of Lake Ohrid underwent seasonal changes from 6.1°C to 27°C in the period 2001 to 2004, temperatures below ~150 m are almost constant around 6°C (Fig. 2a). The top 150 to 200 m water column follows the usual stratification seasonality of deep, temperate lakes. Below ~150 m *in-situ* temperatures show only small depth gradients between –0.02 and 0.15°C per 100 m. The frequently observed temperature increase towards the bottom can be explained partly by isentropic compression with depth leading to an adiabatic lapse rate $\Gamma \approx 0.003$°C per 100 m (Fig. 5a).

Below ~150 m depth the water column is basically stratified by salinity (S) while temperature has but a small effect (Fig. 5b). The salinity-induced stability explains the occurrences of inverse temperature stratification in the hypolimnion, as observed in 2001 and 2003 (Figs. 2a, b, and 5a). S accumulates in the hypolimnion through mineralization as a result of the productive season (see below). Every winter the salinity gradient is partly mixed by increased turbulence. Within the surface layer S is reduced by phytoplankton up-take and calcite formation and increased by evaporation during summer time (Fig. 2b). Dissolved oxygen (DO) shows a similar seasonal pattern as salinity. During stagnant summer season DO is consumed in the hypolimnion and replenished in winter by turbulent mixing and potentially by plunging rivers (Fig. 2c).

Occasional Complete Overturn

The term "complete overturn" will be used in the following to describe deep convective winter mixing which consequently leads to a density destratification of the whole water column. Lake Ohrid is an oligomictic lake, with complete overturn occurring roughly every seventh winter (Hadzisce 1966). In the absence of such an event a gradual increase in temperature of ~0.025°C yr^{-1} was observed in the hypolimnion from 2001 to 2004 (Fig. 6a), corresponding to an average heat input of ~0.12 W m^{-2}. This warming is the result of turbulent vertical heat flux, geothermal heat flux and subaquatic springs.

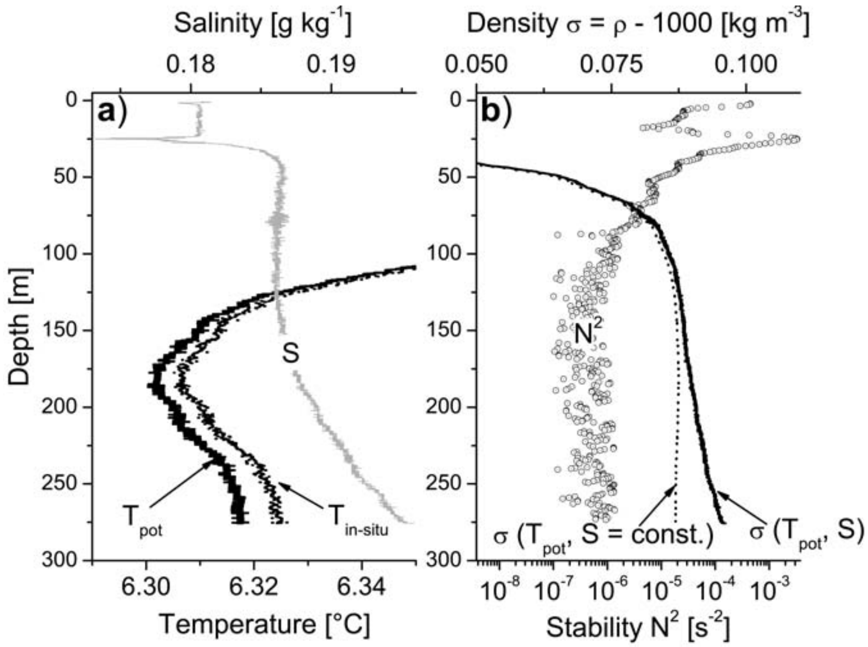

FIG. 5. Stratification, based on a CTD-profile taken on 17 Sep. 2003. (a) Lake salinity, based on measured conductivity, in-situ temperature, as well as potential temperature with reference to lake surface. (b) Calculated profile of potential density, as well as water column stability N^2.

From March to November 2003 the temperature gradient at 200 m depth was almost zero and consequently the turbulent vertical heat flux was negligible. Still, temperatures in the deep water during this period continued to increase, as confirmed by thermistors and CTD profiles. Based on the two thermistors in Figure 6a the heat input of 0.069 W m^{-2} was found. Geophysical surveys for the area of Ohrid indicate a geothermal heat input of ~0.035 W m^{-2} (Cermak and Hurtig 1979, Hurtig *et al.* 1992). Using T ≈ 10°C in surface springs, the remaining increase of 0.034 W m^{-2} could be explained by subaquatic spring inflows below 200 m depth of 0.3 m^3 s^{-1}.

In March 2004 a complete overturn was observed (Figs. 2 and 6), triggered by a two-week period when lake surface temperatures dropped by ~0.1°C to 6.1°C (Fig. 7). Nevertheless the surface temperature in the two preceding winter seasons was only 0.1°C higher than in March 2004 (Fig. 7), illustrating the sensitivity of the lake to surface cooling. To better understand the phenomenon we divided the vertical structure into two layers. A layer MIX above ~200 m, which is mixed with the surface water annually and a bottom layer BOT, which is only mixed during irregular complete overturn. Such a complete overturn can occur if the density in MIX is higher than the density in BOT. In the case of Lake Ohrid the salinity-induced stability must be overcome by negative buoyancy from cooler surface temperatures. This is the case if the so-called mixing ratio

$$0 < R_\rho = \frac{\beta \cdot (S_{BOT} - S_{MIX})}{\alpha \cdot (T_{BOT} - T_{MIX})} < 1 \qquad (1)$$

where $\beta = 0.805 \cdot 10^{-3}$ (g kg^{-1})$^{-1}$ is the coefficient of haline contraction based on Lake Ohrid salt composition, α [°C^{-1}] is the thermal expansivity, S [g kg^{-1}] is salinity, T [°C] is potential temperature. Complete overturn can occur if T_{MIX} becomes less than T_{MIX} at $R_\rho = 1$. From equation (1) we find this theoretical mixing temperature

$$T_{MIX}(R_\rho = 1) = T_{BOT} - \frac{\beta}{\alpha} \cdot (S_{BOT} - S_{MIX}) \qquad (2)$$

Applying equation (2) to data from CTD-profiles and thermistors from 1999 to 2004 yields the function plotted in Figure 7. Surface temperatures

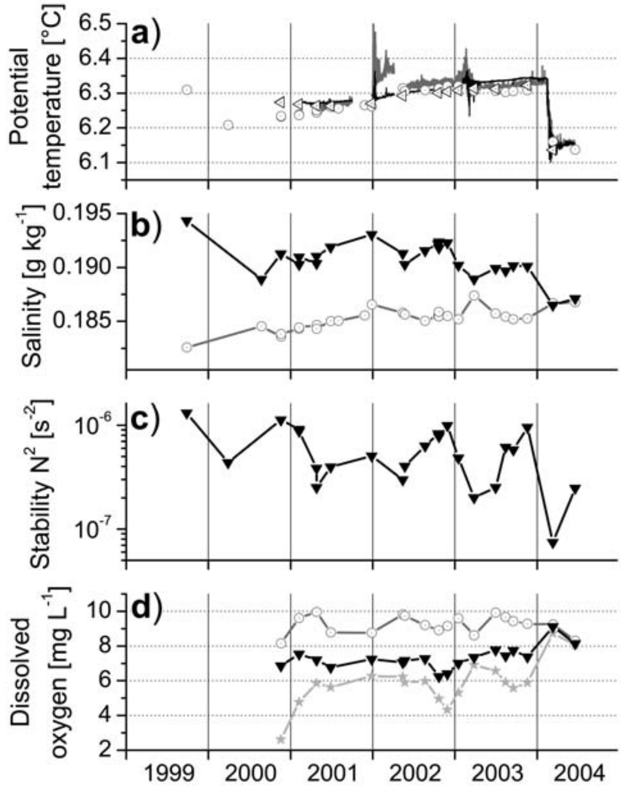

FIG. 6. Temporal development in deep hypolimnion. (a) Change in temperature, from thermistors (lines) and CTD profiles (symbols). Gray line / circles are at 200, black line / triangles are at 277 m depth. The offset between CTD and thermistor measurements is because of smaller accuracy of the CTD probe. (b) Volume-averaged salinity, 0–200 m (gray circles), 200–250 m (black triangles). (c) Average stability of stratification between 200 and 250 meters depth from CTD measurements. (d) Volume-averaged content in dissolved oxygen from CTD profiles and Winkler measurements, 0–200 m (gray circles), 200–250 m (black triangles); DO directly above bottom (light gray stars).

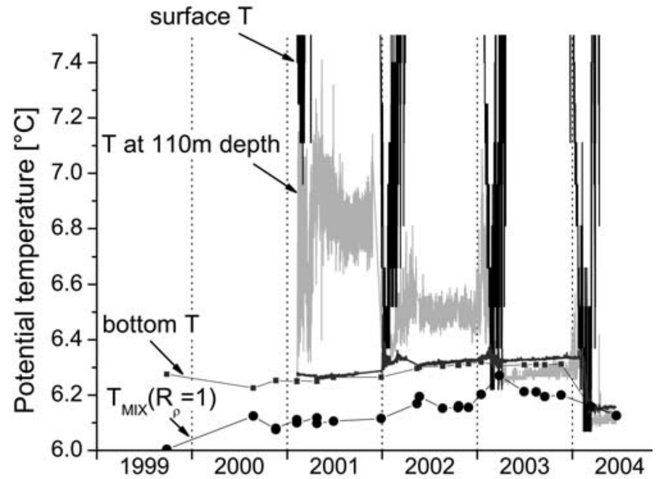

FIG. 7. Water temperature 1999–2004. Temperature at surface and at 110 m depth was measured with Vemco temperature loggers, bottom temperature is based on RBR TR1050 temperature loggers at 200 and 272 m depth (line) and regular Seabird CTD profiles (squares). T_{MIX} (R_ρ = 1) is calculated using equation (2) for one layer above and one below 200 m depth.

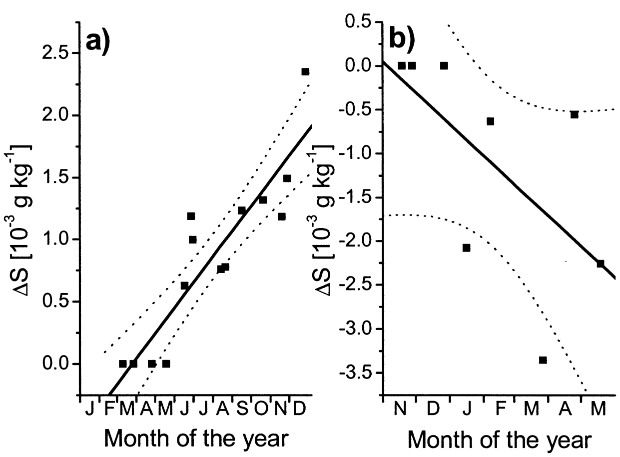

FIG. 8. Salinity dynamics below 200 m depth from CTD profiles covering the period 2000 - 2004, excluding the deep convection in March 2004. Bold lines are linear regressions; dotted lines are respective 95% confidence limits. (a) Average seasonal salt increase from mineralization during summer of $1.8 \cdot 10^{-3}$ g kg^{-1} or $8.44 \cdot 10^3$ metric tons (R = 0.90). (b) Average seasonal salt decrease from enhanced mixing during winter of $-1.05 \cdot 10^{-3}$ g kg^{-1} or $4.92 \cdot 10^3$ metric tons (R = 0.61).

stayed higher than T_{MIX} (R_ρ = 1) until winter 2003/2004, when a complete overturn took place. T_{BOT} increased at a rate of ~0.025°C yr^{-1} until complete overturn.

The bottom layer salinity S_{BOT} depends on the intensity of mineralization, as well as the erosion of salinity stratification in winter (Fig. 6b). The two processes are individually assessed for the period 2000–2004 in Figure 8, excluding the complete overturn in 2004. Annual net mineralization is very

similar over the time of observations (Fig. 8a). The net erosion of salinity during winter shows more fluctuations, as it depends on turbulent mixing by surface cooling and wind forcing (Fig. 8b). Linear regressions in Figure 8 show that on average the expected annual salt increase from mineralization within BOT is about 75% or 4 metric tons higher than the turbulence-induced decrease in winter. While this relation can be reversed in particular winters (e.g., 2003; Fig. 6b), on average the salinity-induced stability increases between events of complete overturn. Equation (2) shows that the increase in S_{BOT} is equivalent to a decrease in T_{MIX} ($R_\rho = 1$) by 0.017°C yr^{-1}. Nevertheless, taking into account both the observed temporal changes in T_{BOT} and S_{BOT} we find a net annual increase in T_{MIX} ($R_\rho = 1$) of 0.025 – 0.017 = 0.008°C yr^{-1}. Thus, every year without complete overturn, the probability for a complete overturn will increase. If we assume a similar future temperature and salinity increase as observed from 2000 to 2003 it would take ~8.0 and 6.6 years, respectively for the deepwater to achieve the T and S levels of December 2003.

Probability of Complete Overturn

Whether a complete overturn occurs depends on the lowest water temperature in winter $T_{min,MIX}$. Based on temperature measurements in Lake Ohrid from 1979 to 2004 (1979–1999, Hydrobiological Institute, unpublished data), performed at least monthly, an average $T_{min,MIX}$ of 6.56°C was found. The observed complete overturn in March 2004 (Fig. 6a) occurred at $T_{min,MIX}$ = 6.10°C, 0.45°C below average. Based on a histogram of the observed T fluctuations in Figure 9 occurrence probability for a specific T deviation can be assessed. As only negative T deviations (cold events) are important for the discussion of complete overturn we used a one-sided histogram for this assessment, assuming normal distribution. A negative deviation of 0.45°C then leads to an expected occurrence probability of 13%. This probability corresponds to a reoccurrence periodicity of complete overturn of 7.8 yr, which consistently supports the reported value of ~7 yr by Hadzisce (1966). Figure 9 compares the data series for $T_{min,MIX}$ with long-term data from Lake Geneva (V = 89 km^3, max. depth = 310 m) and Lake Constance (V = 48 km^3, max. depth = 254 m), two lakes of oligomictic nature with comparable volumes as Lake Ohrid. Surprisingly deviations

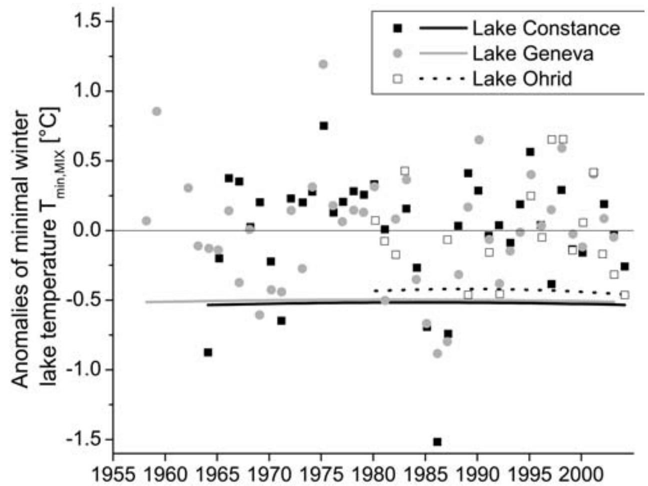

FIG. 9. Variation in minimal lake surface temperatures $T_{min,MIX}$ in winter for Lake Ohrid, Lake Geneva, and Lake Constance. Symbols are based on de-trended measurements from at least monthly temperature profiles and thermistors. Lines are one-sided prediction intervals at 87% level. Data for Lake Constance are from IGKB (International Commission for the Protection of Lake Constance), Langenargen (GER); data for Lake Geneva are from CIPEL (International Commission for the Protection of Lake Geneva), Lausanne (CH).

from average $T_{min,MIX}$ differ by less than 20% among the three lakes (Fig. 9).

Turbulent Mixing

During periods without complete overturn, vertical exchange between MIX and BOT is governed by stratified turbulent mixing. The heat budget method (Powell and Jassby 1974) was used to assess turbulent diffusion coefficients K_z ($m^2\ s^{-1}$) for the summer periods 2002 and 2003. For this particular application of the method, we assumed that the observed heat input below a certain depth is the result of a heat flux from above and the combined geothermal and groundwater heat input of ~0.069 W m^{-2} (see above: occasional complete overturn). In summer an average $K_{z,sum} = 3.5 \cdot 10^{-5}\ m^2\ s^{-1}$ was found at the interface between MIX and BOT in 200 m depth. Based on the net observed reduction in salinity in BOT (Fig. 8b) and the input from River Sateska, the winter diffusivity $K_{z,win}$ was found to be $2.8 \cdot 10^{-4}\ m^2\ s^{-1}$, almost one order of magnitude greater than $K_{z,sum}$. The result of this in-

creased turbulence can also be seen in temperature fluctuations at depths of 200 m and 277 m every winter (Fig. 6a).

The observed mixing is also partly the result of buoyancy flux J_b [W kg^{-1}] caused by the subaquatic springs

$$J_b = \frac{g}{\rho} \cdot \left(\frac{Q \cdot \Delta\rho}{A} \right) \tag{3}$$

where g [m s^{-2}] is acceleration due to gravity, Q [m^3 s^{-1}] is the spring inflow, $\Delta\rho$ [kg m^{-3}] is the density difference from ambient lake water and A [m^2] is the cross sectional area at the considered depth. J_b can be transformed into K_z using the Osborn (1980) relation

$$K_z = \gamma_{mix} \cdot \frac{\varepsilon}{N^2} \sim \frac{\gamma_{mix}}{2} \cdot \frac{J_b}{N^2} \tag{4}$$

where ε [W kg^{-1}] is the dissipation of turbulent kinetic energy equal to 1/2 J_b for a buoyant plume (Wüest and Lorke 2003), N^2 [s^{-2}] is the stability frequency and γ_{mix} [-] is the mixing efficiency, set to 0.2, as typical for convective turbulence (Ivey and Imberger 1991). We applied equations (3) and (4) to the plume model of McGinnis *et al.* (2004), assuming 20 springs with inflows of 0.5 m^3 s^{-1} distributed between 200 and 50 m depth. The result shows that particularly below 100 m depth the contribution to basin-scale vertical diffusivity by subaquatic springs is of the same magnitude as observed $K_{z,sum}$.

Large-scale Wind-induced Mixing

An interesting phenomenon has been first observed by Stankovic and Hadzisce (1953), while profiling temperature on east-west transects in the northern part of the lake. On many instances they found that the isotherms are bent downwards approaching the shore forming a cone-shaped structure. These findings were verified with CTD transects both in N-S and E-W direction in June 2004 (Fig. 10a). Remote sensing images from 2001 to 2004 indicate cooler surface water in the northern lake center on many instances, often extended over time periods of several days. Before and during the measurements presented in Figure 10a remote sensing images also confirm the persistency of the phenomenon (Fig. 10b). Thus the cone-shaped structure cannot be the result of oscillating processes, such as internal seiching. The plausible explanation of such a persistent feature is provided by the so-called "Ekman pumping." In the bounded water body of a stratified lake, wind parallel to the shore causes downwelling/upwelling concentrated along the shore to within a distance, given by the baroclinic Rossby radius a = c f^{-1}, where c is the speed of long internal waves [m s^{-1}] and f [s^{-1}] is the Coriolis parameter (Gill 1982). For Lake Ohrid ~6.5 km is found in June 2004, well within the lake shoreline. The wind predominantly blows along the north-south axis and most often toward the south. Given the protection of the north-eastern lake side by hills, wind toward the south would be expected mainly along the western shore, whereas wind toward the north is probably equally distributed. In combination an anti-clockwise

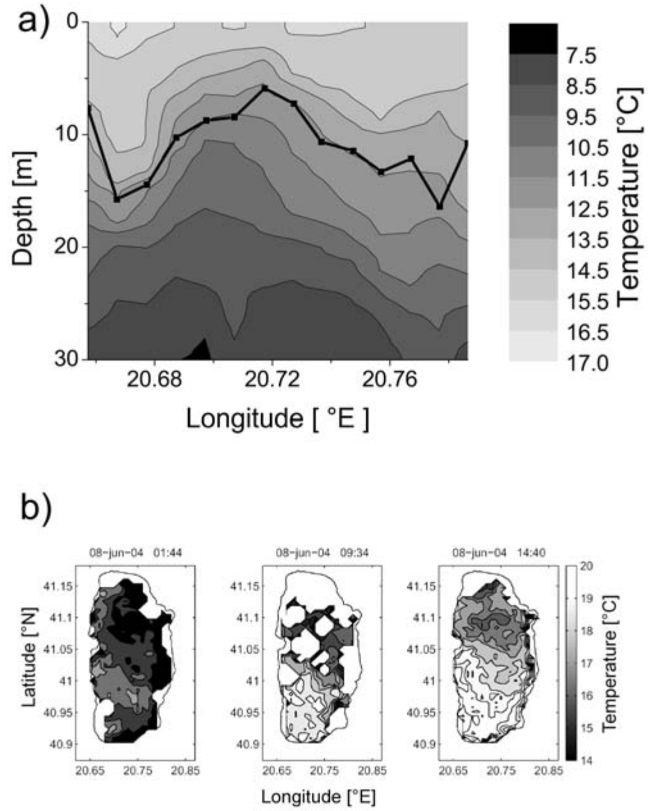

FIG. 10. Upwelling in the northern part of Lake Ohrid on 8 June 2004. (a) Temperature contour from CTD transect at 41.105° N latitude from 10:30 to 14:30. Maximum pH indicates depth of major photosynthetic production (connected symbols). (b) Surface temperatures from remote sensing based on AVHRR measurements from NOAA satellites (detailed method in Oesch et al. 2005). White areas indicate missing values (e.g., because of clouds).

TABLE 3. Oxygen and salt balance of Lake Ohrid in absence of complete deep convection.

Process	Governing Equation[1]	Average annual rates of change		
		DO $[mg\ L^{-1}\ yr^{-1}]$	Salinity $[g\ kg^{-1}\ yr^{-1}]$	
		200–250 m (BOT)	0–200 m (MIX)	200–250 m (BOT)
River input (to BOT when plunging)	$\frac{Q_{riv,i}}{V_i} \cdot (X_{riv} - X_{LO,i})$	0.05	$0.56 \cdot 10^{-3}$	$1.21 \cdot 10^{-3}$
Spring input (surface and subaquatic)	$\frac{Q_{spring,i}}{V_i} \cdot (X_{spring} - X_{LO,i})$	0.006	$0.80 \cdot 10^{-3}$	$0.12 \cdot 10^{-3}$
Turbulent diffusion in winter (during 2 months)	$K_{z,win} \cdot \frac{\Delta X}{\Delta z} \cdot \frac{A_{200m}}{V_i}$	0.81	$0.21 \cdot 10^{-3}$	$-2.86 \cdot 10^{-3}$
Turbulent diffusion in summer (during 10 months)	$K_{z,win} \cdot \frac{\Delta X}{\Delta z} \cdot \frac{A_{200m}}{V_i}$	0.50	$0.13 \cdot 10^{-3}$	$-1.40 \cdot 10^{-3}$
Evaporation (salinity increase at lake surface)	$\frac{Q_{evap}}{V_{MIX}} \cdot S_{MIX}$	—	$1.49 \cdot 10^{-3}$	—
Salt loss to sedimentation	constant	—	$-0.34 \cdot 10^{-3}$	—
Outflow	$-\frac{Q_{out}}{V_{MIX}} \cdot S_{MIX}$	—	$-2.84 \cdot 10^{-3}$	—
Mineralization	constant	–1.67	-	$3.09 \cdot 10^{-3}$
Total inputs		1.37	$3.21 \cdot 10^{-3}$	$4.41 \cdot 10^{-3}$
Total outputs		–1.67	$-3.21 \cdot 10^{-3}$	$-3.66 \cdot 10^{-3}$
Annual change		–0.30	0	$0.75 \cdot 10^{-3}$

[1] Rates of change are in $[g\ kg^{-1}\ s^{-1}]$ for S and $[mg\ L^{-1}\ s^{-1}]$ for DO, X is either DO $[mg\ L^{-1}]$ or S $[g\ kg^{-1}]$, subscript i refers either to compartment MIX or BOT, subscript LO stands for Lake Ohrid, Q $[m^3\ s^{-1}]$ are average annual flow rates, V $[m^3]$ is compartment volume ($V_{MIX} = 50.3 \cdot 10^9\ m^3$ and $V_{BOT} = 4.7 \cdot 10^9\ m^3$), $A_{200m} = 149.8 \cdot 10^6\ m^2$ is cross-sectional lake area at 200 m depth, $K_{z,win} = 2.8 \cdot 10^{-4}\ m^2\ s^{-1}$ and $K_{z,sum} = 3.5 \cdot 10^{-5}\ m^2\ s^{-1}$ are turbulent diffusion coefficients, $\Delta z = 125$ m is depth difference between MIX and BOT.

current seems to evolve along the northern lake shore with downwelling at the shore from Coriolis force and finally an up-welling in the center. The downwelling velocity of Ekman pumping can be described in a two-layer approach by $c \cdot \tau \cdot (\rho \cdot g' \cdot H_1)^{-1}$ (Gill 1982), where τ $[N\ m^{-2}]$ is the wind stress, $g'[m\ s^{-2}]$ is the reduced gravity ($= g \cdot \Delta\rho/\rho$) and H_1 is the depth of the surface layer. As wind speed exceeds 6 m s^{-1} only during ~3% of the time in summer, downwelling velocities are small (wind data for station in Ohrid, U.S. National Climatic Data Center, Station Nr. 135780). Using the above term it would take about 20 hours of a wind speed of 5 m s^{-1} to create a downwelling of surface water to 15 m depth as seen in Figure 10a. If indeed Ekman pumping is the reason for the observed temperature gradients both south and north winds have to contribute to an anticlockwise current.

Oxygen and Salinity Balances

The processes described above influence the mass balances of DO and S (Table 3). As defined above, the water column is divided into an annually mixed section MIX above 200 m depth and the deep hypolimnion BOT. To close the balance for S_{MIX} we set $\Delta S_{MIX}/\Delta t = 0$, as the major fluxes are external and can be expected to be in equilibrium. The balance for S_{BOT} is based on observed net mineralization and net winter loss (Fig. 8a and b). For DO_{BOT} it is assumed that the observed difference between MIX and BOT (Fig. 6d) is built up during ~8 years of permanent stratification, resulting in an average annual net decrease $\Delta DO_{BOT}/\Delta t = -0.3$ mg L^{-1} yr^{-1}. DO_{MIX} is returned to 89% saturation annually during winter convection.

S_{BOT} and DO_{BOT} are governed by diffusive exchange with MIX and mineralization of settled material. Plunging of River Sateska contributes about one fourth of the salinity input to BOT, whereas it is almost negligible for DO input for the current situation. However its effect will increase with growing difference to the ambient lake concentration (equation in Table 3). For S_{BOT} the calculations are based on observations in Figure 8a and b. For DO_{BOT} the calculated summer net decrease of 1.16 mg L^{-1} yr^{-1} (Table 3) is in agreement with observed 0.79 ± 0.28 mg L^{-1} yr^{-1} from 2000–2004. If the annual fluxes (except diffusion) are constant DO_{BOT} will approach a level, where diffusion equals consumption. Under the above assumptions this "equilibrium" DO level is found at 6.7 mg L^{-1}. Indeed the observed volume averaged DO between 200 and 250 m is above this value (Figure 6d). Nevertheless the DO level directly above the lake bottom can be up to 4 mg L^{-1} lower than the average DO_{BOT} (Fig. 6d). Evaporation and the main outflow are dominating the salt balance for S_{MIX}, followed by spring input. Net salt loss from MIX through adsorption to settling particles is comparably small. Given the numbers in Table 3, 80% of the salt leaving MIX with settling particles is expected to be mineralized in BOT, significantly more than the 13 and 60%, found by Ramisch *et al.* (1999) for two eutrophic lakes. This difference can be explained by the relatively high contents in dissolved $CaCO_3$ in the lakes observed by Ramisch *et al.* (1999). Still it seems likely that we underestimate the salt loss to sedimentation. Such an underestimation could be explained by the high sensitivity of the salt loss term to errors in the balance: e.g., even an increase of 10% in evaporation would increase the sedimentation term to the expected range (~60% re-dissolution). However, given the small effect of the unknown error in the salt balance in MIX, no adaptation has been made.

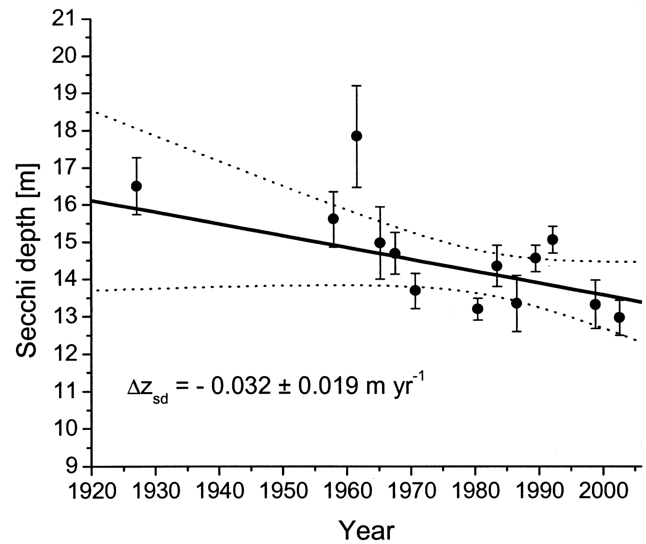

FIG. 11. Pelagial Secchi depth measurements. Circles indicate 3-year averages from productive season (Mar–Oct), error bars represent error of mean. The bold line shows linear regression using (errors)$^{-2}$ as weight, dotted lines are 95% confidence limits. Data are taken from Stankovic (1960), Ocevski (1974), unpublished data from Hydrobiological Institute (1979–1995), Patceva (2001) and Patceva (2005).

Water Clarity

Lake Ohrid is an exceptionally clear lake with an average Secchi depth z_{sd} of ~14 m (Fig. 11). The clarity was also confirmed in a PAR-profile (400 to 700 nm) taken in June 2004 using a pair of spherical LI-COR sensors. In the profile the visible light penetrated below 90 m with a 1‰-compensation depth (Kalff 2002) between 50 and 60 m and an average extinction coefficient of $\eta = 0.13 \pm 0.004$ m^{-1}. This value is comparable to values by Kalff (2002) from clear and oligotrophic lakes, such as Lake Perry (0.2 m^{-1}) and Crater Lake (0.05 m^{-1}). The empirical relation from Wetzel (2002)

$$\eta = 1.58 \cdot z_{sd}^{-1} \qquad (5)$$

yields $\eta = 0.11$ m^{-1}, so that the Secchi and PAR estimates of η are in agreement.

TABLE 4. Human impacts under historic, current, and future scenario.

Human impact	Unit	Historic situation[1] A	Status quo[1] B	Future development[1] C1	C2
***Eutrophication**:*					
TP concentration in Lake Ohrid	mg P m^{-3}	2.25	4.5	9	
Changes in water balance:[1]					
Annual inflow					
—surface springs	$m^3 s^{-1}$	13.6	10.3	2.9	
—subaquatic springs	$m^3 s^{-1}$	10.8	9.9	6.6	
—River Sateska	$m^3 s^{-1}$	—	5.5	3.6	
—other tributaries	$m^3 s^{-1}$	3.9	3.4	3.2	
Total change from historic (A)	$m^3 s^{-1}$	—	+0.8	–12	
Global warming:					
Increase of lake surface T	°C yr^{-1}	0	0.006	0.04	0
Increase in total lake temperature (new equilibrium)	°C	—	—	—	4

[1]Detailed account of the scenarios is given in the text
[2]Status quo assumes 2 $m^3 s^{-1}$ use of spring water and 1 $m^3 s^{-1}$ from other tributaries during summer season. Runoff measurements are based on own measurements (2002–2003) and data from Ivanova (1974), the Macedonian Hydrometeorological Institute (unpublished data) and the Hydrobiological Institute Ohrid (unpublished data).

The clear water reflects oligotrophic conditions, with a production maximum often below z_{sd} and with generally low phytoplankton concentrations (average chlorophyll A (ChlA) during the productive periods 2001–2003 of ~1.3 ± 0.1 mg m^{-3} (Patceva 2005)). Moreover the "filtered" spring water and the relatively small surface runoff lead to exceptionally small loads of suspended particles.

DEFINITION OF HUMAN IMPACTS

In the following observations of anthropogenic changes in the physical boundary conditions are presented and their future potential discussed. For the impact assessment the changes are grouped into four scenarios:

(A) the historic situation, about one century ago, before major human impacts occurred,
(B) the observed status quo,
(C1) expected future development during global warming process, and
(C2) at a new, warmer temperature equilibrium.

The main assumptions for each scenario are summarized in Table 4.

Eutrophication

As mentioned above, phosphorus (P) is the key eutrophication factor in Lake Ohrid. This eutrophication is mainly the result of increased human population in the catchment area and probably progressive use of P-containing washing agents and of fertilizer in agriculture. However, eutrophication is hard to track, because of the low P-concentrations (Total phosphorus (TP) ~4.5 mg m^{-3}) and the short history of precise TP-measurements. Still, clear changes are visible in the close vicinity of polluted inflows, with increased nutrient concentration, shifts in species composition and appearance of coliforms (Watzin *et al.* 2002, Patceva *et al.* 2004, Lokoska and Jordanoski 2004). Combining the observed increase in sedimentary P (Fig. 12) with inflow measurements in a simple linear phosphorus model (Vollenweider 1969), revealed a ~3 fold increase in TP-concentration since historic times and a doubling over the past ~100 yr (Matzinger *et al.* 2004). In the following assessment we assume 50% reduced TP concentration in the lake for scenario (A) and a further doubling in the future for (C) (Table 4). The assumption of slow increase in scenario (C) seems plausible, given the potential ongoing population increase, intensified agriculture but concurrent improvement of sewage treatment (Ernst Basler and Partners 1995).

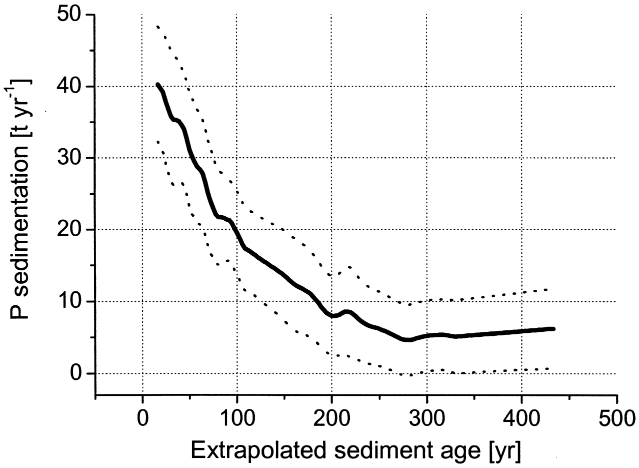

FIG. 12. Bold line is average sedimentation of TP, dotted lines indicate error of mean. Results are based on three sediment cores taken along the north-south axis at 122, 202, and 282 m depth in 2002 and 2003. Sediment age is based on ^{210}Pb measurements and sediment trap results from 2001–2004.

Changes in Water Balance

As the area is rather dry with hot summers, water is generally scarce, particularly for use in agriculture. During the summer season, water is diverted from all tributaries of both lakes for irrigation, reducing their flow rate (e.g., ΔQ ~2.5 $m^3\ s^{-1}$ for River Sateska) or setting them completely dry. Moreover there are groundwater pumps and systems to take water directly from upstream Lake Prespa. The growing need for irrigation water has its most obvious effect on Lake Prespa, where the water level has dropped by ~6 m over the past decade (Matzinger *et al.* 2006). The water use today reduces the annual natural water inflows by 4-5 $m^3\ s^{-1}$, excluding River Sateska. This estimated reduction is composed of 40% due to irrigation in the Lake Prespa catchment (Matzinger *et al.* 2005), 40% due to use of surface spring water (mainly for water supply but also for irrigation in Albania), and 20% by irrigation in the north of Lake Ohrid. The historic scenario (A) is simplified by assuming that then no significant water use was present.

Apart from the reduction in water inflow, another major change in status quo (B) compared to (A) is the artificial diversion of River Sateska in 1962, which contributes about 5.5 $m^3\ s^{-1}$ to the total river inflow today and thus compensates for the recent water losses described above. (Table 2, Fig. 1).

In a worst case future scenario (C) the tributaries and the surface spring inflows could be used completely during the summer season from May to October. Moreover Lake Prespa could basically be set dry. Taking into account the summer runoff of the rivers and the surface springs from Lake Prespa (Matzinger *et al.* 2005), the annual water balance would be reduced by 13 $m^3\ s^{-1}$ compared to status quo. During the summer season the surface springs and the tributaries would be used to their full capacity of 10.3 and 4 $m^3\ s^{-1}$, respectively and the subaquatic springs would be reduced by their current Lake Prespa share of 3.3 $m^3\ s^{-1}$ (Table 4).

Global Warming

CO_2 levels in the atmosphere are very likely to increase in the next century (IPCC 2001). The potential effects of two IPCC scenarios A2 and B2 (CO_2 level increase from the base level of 353 ppm to 822 and 1,143 ppm, respectively, in the year 2100) on European climate have been simulated with coupled global and regional circulation models by Räisänen *et al.* (2004) and Giorgi *et al.* (2004) who predict a significant increase in air temperature for the southern Balkan Peninsula from 1961–1990 to 2071–2100 of 4 ± 2°C and 3.6 ± 0.9°C, respectively, substantially larger than the 0.6°C increase observed over the 20th century (IPCC 2001). It was shown that recent changes in air temperatures are also well reflected in average lake temperatures, whereas a smaller temperature increase is expected in the lakes' deep water, due to limited exchange with the main water body (cf., Livingstone 2003, O'Reilly *et al.* 2003). Observations for the deep water of Lake Ohrid since the 1920s show an increase of 0.005 ± 0.001°C yr^{-1} ($R = 0.37$, $p < 0.0001$) (Fig. 13).

In the following discussion we will use the expected increase in air temperature of 0.04°C yr^{-1} for scenario (C), 0.006°C yr^{-1} for status quo (B) and no increase for (A). We further assume that the potential increase in air temperature is directly transferable to surface water. This assumption is confirmed by observations (Livingstone and Lotter 1998) and the coupling of global circulation models with hydrological models (Mortsch and Quinn 1996, Blenckner *et al.* 2002). Based on the model results of Giorgi *et al.* (2004), the magnitude of interannual variation in winter air temperatures is assumed to be unaffected by climate change. Changes in other climate parameters, such as wind speed or precipitation, are not considered in our analysis, because their simulated changes were often in the

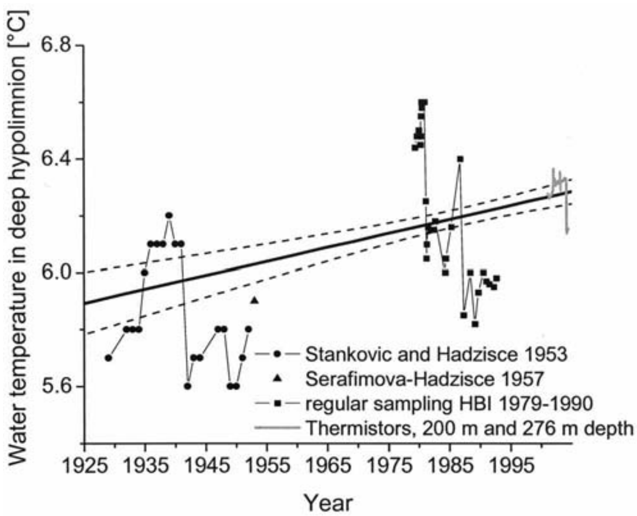

FIG. 13. ***$T_{in\text{-}situ}$ measurements from different sources in the deep water between 200 and 285 m depth. The bold line is linear regression; dotted lines are 95% confidence limits.***

range of expected variability (Räisänen *et al.* 2004, Giorgi *et al.* 2004).

In assessing the impact of an increase in lake surface temperatures we differentiate between (i) the ramp-up period of ongoing warming as expected over the next century and (ii) a new equilibrium with sub-scenarios (C1) and (C2) (Table 4).

DISCUSSION—ANALYSIS OF EXPECTED CHANGES

SRP Supply between 20 and 150 m Depth

As has been shown in the approach endemic phytoplankton species are dependent on nutrient supply, particularly bio-available SRP, to depths between 20 and 150 m. This nutrient supply is based on the input of springs, rivers and mineralization of organic matter at the lake bottom, as well as Ekman pumping in the upper water column. It can be expected that the water circulation of the top 20 m by Ekman pumping would remain unchanged under our assumption of constant wind speeds in all scenarios. However inflow and mineralization are affected by changes in the water balance and eutrophication.

The main source of nutrients come from surface and subaquatic springs, which continuously supply SRP to the 20–150 m depth interval throughout the year (Fig. 4). Based on the assessment of actual and historic P concentrations in Lake Prespa and precipitation-fed springs by Matzinger *et al.* (2006) the spring P-loads for the three scenarios can be calculated. It was found that the input during 6 summer months increased from scenario (A) to scenario (B) by merely 2% to the current load of ~$1.8 \cdot 10^3$ kg P. This increase, despite a reduction in spring inflow (Table 4), is the result of observed ongoing eutrophication of Lake Prespa (Matzinger *et al.* 2006). However, under scenario (C), without inflow from Lake Prespa and no surface spring inflow in summer, the spring P-input would be reduced to ~$0.3 \cdot 10^3$ kg P, ~16% of the current value. This dramatic reduction contrasts the overall eutrophication from (A) to (C) (Table 4).

TP-sedimentation in the past has developed in parallel to lake P-concentration (Fig. 12). This proportionality is likely to be true into the future, both for TP-sedimentation and P-concentration of River Sateska (P-concentrations taken from Jordanoski 1999). Based on SRP measurements in the water column it is assumed that ~50% of the settled TP is released as SRP and makes its way upward through the water column (Matzinger *et al.* 2004). Based on these assumptions a maximal annual SRP-flux from the lake bottom of 10, 20, and $40 \cdot 10^3$ kg P yr^{-1} is expected for scenarios (A), (B), and (C), respectively, much higher than contributions from springs. Summarizing the above we found that

(i) there is a significant increase in phosphorus availability from scenarios (A) through (C) close to the lake surface and toward the bottom from anthropogenic eutrophication, but
(ii) up to an 80% decrease in continuous SRP supply between 20 and 150 m depth during the summer season must be expected from reduced spring inflow.

Water Clarity in the Trophogenic Zone

Water clarity is certainly a crucial factor for the endemic phytoplankton species, given their deep habitat with light availabilities around 1 ‰ of surface radiation. Given the very low inorganic particle concentrations in Lake Ohrid, water transparency depends mainly on phytoplankton. Contrary to the natural nutrient input, which stems predominantly from the karst aquifers and plunges into the thermocline or below during the productive season, the bulk of the anthropogenic nutrient sources enter the surface layer of the lake through rivers, sewage channels, and diffuse sources. In

summer, when the rivers are almost dry from upstream irrigation, they still serve as sewage channels from the major towns and villages close to the lake shore (Veljanoska-Sarafiloska 2002). Thus eutrophication would mainly trigger phytoplankton production in the surface layer and reduce light availability at greater depths. For the change from historic (A) to status quo (B), Secchi depth observations indeed show a significant reduction by ~0.05 m yr^{-1} ($p = 1.04 \cdot 10^{-4}$) since 1926 for the summer periods (Fig. 11). For an estimate of the effect of increased productivity we used the empirical relationship from Kalff (2002)

$$\mathrm{ChlA} = 0.407 \cdot \mathrm{TP}^{0.874} \tag{6}$$

where ChlA and TP are to be inserted in [mg m^{-3}]. Equation (6) resulted in a historic ChlA concentration of 0.8 mg m^{-3}, compared to the present of 1.5 mg m^{-3}, which is very close to the measured value of 1.3 mg m^{-3}. The effect of higher ChlA on light extinction coefficient η is $\Delta\eta \approx 0.019$ m^{-1} (mg-ChlA $m^{-3})^{-1}$ (average of values from Smith and Baker 1978 and Megard *et al.* 1980). In our case, we then find an increase in η from 0.117 to 0.13 m^{-1} since 1926, signifying a decrease of the theoretical 1 ‰-compensation depth by ~6 m. In both cases the major share of light extinction is from pure water. Using equation (5) we expect a concurrent reduction in Secchi depth of 1.4 m, which is about 60% of what is observed (Fig. 11). For scenario (C) with doubled TP-level (Table 4) we expect a ChlA level of 2.8 mg m^{-3}, a coefficient η of 0.154 m^{-1}, and finally a further decrease in the 1‰-compensation depth from today by ~8 m. In total a decrease in theoretical 1‰-compensation depth of ~14 m is expected from scenario (A) to (C).

Given the predominance of green algae and cyanobacteria in the top 20 m of the lake (Patceva 2005), these species would also be the main beneficiaries of additional nutrient inputs at the surface. Thus while the overall lake productivity increases, the habitat of deep living, endemic phytoplankton species would be reduced. The negative impact on these communities might even be amplified by their reduced competitiveness under nutrient-rich conditions, as observed in the vicinity of polluted inflows (Watzin *et al.* 2002).

Oxygen Supply to the Hypolimnion

Availability of DO was identified as one main factor for preserving the endemic bottom fauna. The DO content in the irregularly mixed bottom layer BOT is replenished by (i) the plunging River Sateska, (ii) input from subaquatic springs, (iii) diffusive input from MIX, and (iv) complete overturn. On the other hand it is consumed by mineralization of settled organic material (Table 3). Given this multitude of influences each scenario (past, present, future) is assessed separately. Our general approach starts at the observed situation after the complete overturn in June 2004 (Fig. 6), again dividing the lake into two layers MIX and BOT. For each year without complete overturn changes are evaluated iteratively, based on equations in Table 3. Turbulent diffusion coefficients $K_{z,win}$ and $K_{z,sum}$ are re-calculated after each time step based on changes in stability N^2 using equation (4). Finally, changes in initial values allows re-calculation of the theoretical mixing temperature T_{MIX} ($R\rho = 1$) and its deviation from T_{MIX}. Occurrence probability p_{co} of this T deviation can be calculated as discussed above, based here on the combined histogram of observed fluctuations in the minimal winter temperatures $T_{min,MIX}$ from Lake Ohrid, Lake Constance, and Lake Geneva (Fig. 9). In other words, p_{co} is the probability that T_{MIX} ($R\rho = 1$) is reached and thus complete overturn takes place. While the evaluation of S_{MIX}, S_{BOT}, and DO_{BOT} for successive time steps is straight forward, DO_{MIX}, T_{MIX}, and T_{BOT} require a different approach. During winter mixing DO_{MIX} and T_{MIX} adapt to conditions at the surface. In the iterative evaluation T_{MIX} is assumed to reach an average minimal water temperature $T_{min,MIX}$ every winter. This average is set constant at $T_{min,MIX}$ = 6.56°C in the absence of global warming. Based on observations DO_{MIX} is set to 89% of saturation at T_{MIX} every year. Change in T_{BOT} can then be estimated based on observed increase from geothermal and subaquatic springs (~0.014°C yr^{-1}) and turbulent diffusive heat flux from MIX. For the years 2001–2004 this heat flux from MIX to BOT was governed by local positive and negative vertical temperature gradients, which are often observed to be alternating below 150 m depth (Figs. 2a and 6a). As a result our approach using global vertical temperature gradients between T_{MIX} and T_{BOT} overestimates the heat flux by a factor of ~4 during the years of observation. For the model it was assumed that this factor remains the same for varying temperature differences between T_{MIX} and T_{BOT}. Figure 14 shows the results of this iterative approach.

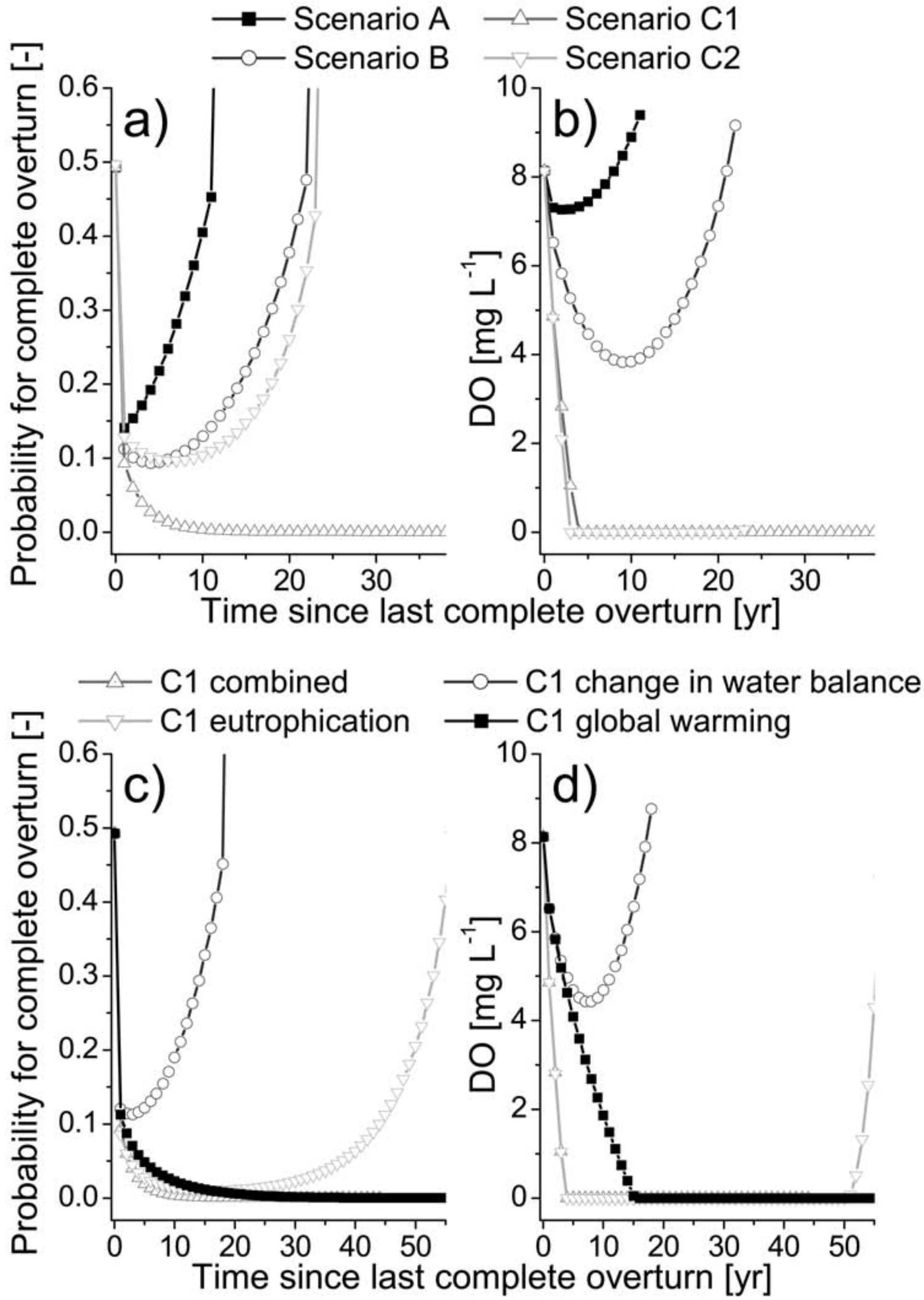

FIG. 14. Probability of complete overturn and DO concentration in bottom layer BOT. (a) and (b) Effect of the different scenarios in Table 4. (c) and (d) Comparison of combined and single effects of anticipated future changes in eutrophication, water use and air temperature. Note the different time scale in (c) and (d).

Scenario (B), Present Status Quo

After winter 2003/2004 p_{co} decreases, as a result of increase in S_{BOT} (Fig. 14a). After 5 yr, rising diffusive salt losses and smaller temperature gradient lead to a decrease in stability and p_{co}. Then, with T_{BOT} getting closer to T_{MIX} complete overturn gets more and more probable until after 23 yr it would occur even at average $T_{min,MIX}$. Similar to salinity, DO consumption from mineralization is outweighed by diffusive input after 9 yr (Fig. 14b). Thus no anoxia is expected under the status quo, even if several warm winters follow each other.

Scenario (A), Historic Conditions

Under historic conditions (Table 4) it was assumed that the 50% reduction in TP concentration is directly transferable to DO consumption, S depletion in MIX, and S accumulation in BOT. Compared to status quo the salinity gradient between S_{BOT} and S_{MIX} stays much lower, because of the lack in plunging River Sateska and the smaller input from mineralization. As a result complete overturn would be expected to have occurred at much shorter intervals 100 yr ago (Fig. 14a). With increasing diffusive flux and reduced DO consumption, DO_{BOT} is only decreasing during the first 2 years after complete overturn (Fig. 14b).

Scenario (C1), Future Conditions during Global Warming

As in scenario (A) the doubling in TP was transferred into a doubling in DO consumption, S depletion in the epilimnion, and S accumulation in the hypolimnion. The raised salt input to BOT from eutrophication together with an annual increase in T_{MIX} from global warming higher than in T_{BOT} from geothermal sources and diffusion dramatically increases the stability of BOT. If no complete overturn occurs in the first few years BOT will basically turn permanently stratified, as long as the situation stays as outlined in Table 4 (Fig. 14a). Because of increased DO consumption DO_{BOT} would turn anoxic after 4 years (Fig. 14b). Figures 14c and 14d show the effect of the three human impacts separately. Changes in water balance had almost no effect on mixing periodicity and DO availability, as the inflows are small relative to the large lake volume (Table 1). A larger impact on DO_{BOT} and S_{BOT} could only be expected from plunging winter inflow of River Sateska, which remains unchanged from scenario (B) to (C). In a separate run for scenario (B) it was found that the diversion of River Sateska reduces p_{co} by ~7%, because of increased S_{BOT}. Despite its input of DO to BOT the stabilizing effect dominates and leads to a reduction of DO_{BOT} by ~1 mg L^{-1}.

Eutrophication is the main reason for smaller DO_{BOT} in scenario (B) and for fast anoxia observed in scenario (C1) (Fig. 14d). Moreover, it supports stratification of the water column through increased mineralization, constraining complete overturn for 35 yr (Fig. 14c). Because of the low stability after 50 yr, BOT could even turn oxic again in the absence of complete overturn (Figs. 14c and d). However accumulation of organic matter, which is not considered in our simplified model, would further delay this process.

Global warming is the process that could ultimately lead to meromictic conditions due to rapid increase in stability of the water column. Even under current DO mineralization BOT would still turn anoxic after 16 years, because of reduced turbulent diffusive DO supply from MIX (Fig. 14d). Under increasingly stable conditions the anoxic layer BOT is also expected to grow in thickness.

Scenario (C2), Future Conditions at Higher Temperature

While the expected warming over the next century would reduce the mixing of BOT dramatically it is interesting to note that the situation is completely different if temperature is assumed to be in a new quasi-steady-state equilibrium, 4°C higher than today (Table 4). Although higher eutrophication supports stratification, p_{co} in (C2) is very similar to (B) (Fig. 14a) and clearly reduces the effect of increased eutrophication (Fig. 14c). The reason lies in the thermal expansivity α, which increases by a factor of 2.3 under an increase in water temperatures from 6 to 10°C at 200 m depth for Lake Ohrid salinity. An increase in α leads in turn to higher T_{MIX} ($R\rho = 1$) (Equation 2) and thus renders complete overturn easier. Still anoxia would be reached fastest of all scenarios because of smaller DO solubility with increased temperature in MIX and in plunging River Sateska.

CONCLUSIONS

Current and potential future human activities will have a major impact on the physical characteristics of Lake Ohrid, which in turn will likely affect the renowned endemism of this unique ecosystem.

Endemic Phytoplankton Species

Deep-growing endemic phytoplankton species require high water clarity and sufficient nutrient supply between 20 and 150 m depth. Future water abstraction from surface spring inflows and Lake Prespa would reduce the continuous nutrient input to this depth interval to ~16% of today's value. However expected eutrophication could compensate for this loss through generally increased phosphorus (P) level and raised release of soluble reactive P from the lake bottom. A much more serious concern is the reduction in water clarity in the surface layers, which is already apparent in trends of Secchi depth measurements. Further eutrophication could lead to a significant decrease in compensation depth by ~14 m relative to historic conditions.

Bottom Dwellers

Endemic bottom dwellers depend upon adequate availability of dissolved oxygen (DO) and thus reach very deep today. While water abstraction has little effect on the overall DO balance, the diversion of River Sateska into Lake Ohrid does affect the lake by the mechanism of deep plunging in winter. However, the resulting salinity input decreases the downward diffusive DO flux into the bottom layer to a larger extent than it replenishes DO by direct transport.

Eutrophication stabilizes the bottom layer through mineralization and increases DO consumption at the sediment. Even a comparably low total P (TP) concentration of 9 mg m^{-3} would extend periods without complete overturn and lead to anoxic conditions in the bottom layer.

A regular increase in temperature from global warming by 0.04°C yr^{-1} would create an almost completely secluded, permanently stratified bottom layer. This layer will turn anoxic as the exchange with the regularly mixed part of the lake decreases. While this tendency towards lower mixing is in line with findings of other authors the situation at a new 4°C higher temperature equilibrium is not. Because of an increase in thermal expansivity with temperature even a more frequent complete overturn would be expected.

Local versus Global Human Impacts

Our results show that both local eutrophication and global warming are jeopardizing the endemic species of Lake Ohrid. Indeed, the effects of eutrophication is already apparent. Moreover, the increase to 9 mg TP m^{-3} is far below the concentrations of most central European lakes, and thus it is a conservative estimate. Since eutrophication can be controlled at a local level it is certainly a priority task for lake management, particularly since its negative effects would be intensified by potential global warming.

However, a more detailed, bio-geochemical assessment is necessary to quantify sustainable P loads under different climate scenarios. The results of this study form an excellent basis for such an assessment regarding the physical lake properties and the sensitivity of Lake Ohrid to human impacts.

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