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TEMPORAL AND SPATIAL SCALES OF LAKE PROCESSES. PART 1: PHYSICAL AND ECOLOGICAL SCALES, NON-DIMENSIONAL PARAMETERS AND FLOW REGIMES

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ABSTRACT
Scaling concepts, related to basic physical transport and mixing processes and other ecological aspects in lakes, are reviewed in an ecohydrodynamic perspective; the length and time scales as well as the non-dimensional parameters governing the various flow regimes and relevant to these processes are also presented, emphasizing the influence they have on lake ecosystems. The hierarchy of length scales is very important in determining both the turbulence structure, and in affecting the ecological processes occurring in lakes. Most of these scales are defined in terms of several basic parameters which, when combined properly, yield important dimensionless numbers, as: R_f, Ri, Pr, T, Re_t, Fr_t, Gr_t, and Pr_t, the flux and gradient Richardson numbers, the Prandtl number, a dimensionless time, and the turbulent Reynolds, Froude, Grasshof and Prandtl numbers, respectively. A few more dimensionless numbers (as the Lake and Wedderburn numbers, L_N, W, etc.) are also useful in characterizing the general dynamic behavior of lakes. Ratios of length or time scales may also provide similar useful parameters.

KEYWORDS: Lake processes, transport, mixing, length scales, time scales, hierarchy of scales, dimensionless numbers, dynamical regimes.

1. INTRODUCTION

1.1 Generalities

There is an increasing need to understand the behavior of lake ecosystems and their dynamic response to external forcing, whether the latter refers to meteorological conditions, inflows and outflows, a source of localized pollution or a climate change. To protect lake environments from further ecological decline, detailed information on a wide range of processes (occurring in the lake body and affecting the quality of water) is needed in order to validate proper water quality management models and evaluate alternative strategies for maintaining and/or restoring lake trophic conditions to an acceptable level.

Lakes are exposed to several disturbing influences. First, the meteorological conditions in the area determine (through wind stirring, radiation, convection and evaporation) the momentum, heat and mass exchanges across the lake's air-water interface (Curry and Webster, 1999). Second, the water from any inflowing streams may impart kinetic and potential energy to the lake water, and third, if outflow exists, some of the energy of the outflowing water may be transformed into kinetic energy of the lake's water. The influence of the earth's rotation may also become significant, if the lake is large. The response to these disturbing influences, constrained by the stratification and bathymetry of the lake, produces an ensemble of motions which may conveniently be divided into three classes: a) basin scale motions (currents-seiching), b) surface and internal waves, eddies (i.e. non-symmetric waves) and intrusions, and c) small scale turbulent motions responsible for intimate mixing.

Data obtained from several lakes has allowed various investigators to postulate recently a, perhaps, more precise picture of the motions developed in a stratified lake, under the influence of the disturbances described previously. Initially, the wind transfers momentum (and energy) to the surface layer which, then, induces seiching of the thermal structure; the initial motion, thus, generated may be decomposed into various modes of seiching, Kelvin and Poincare waves and forced gravity waves. Much of this energy is transferred, internally, into higher frequency internal non-linear solitary waves, bores and hydraulic jumps through non-linear steepening and shoaling where the metalimnion intersects the lake boundary, forming a spectrum of internal wave motions, in the metalimnion, with frequencies from about $10^{-6} \div 10^{-2}$ Hz depending on the size of the lake and the severity of its stratification. The higher frequency internal waves propagate, in groups, throughout the lake and their energy is absorbed when they impinge on the lake boundary.

The hydrodynamics of motions developed in a lake expresses a delicate interplay among the four dominant disturbances (wind, inflow, outflow, differential heating), the potential energy of the resident stratification, the bathymetry of the lake and the earth's rotation (if the lake is large); for small to medium size lakes it (the hydrodynamics, that is) depends on the stratification of the lake and the instabilities developed. The disturbing forces work against the potential energy gradient set up, primarily, by heat radiation. Thus, the momentum imparted by the wind to the lake leads to large basin scale motions in the form of Kelvin or Poincare waves or gravitational seiching, which degenerate into smaller scale internal waves and intrusions ultimately used by a host of instability mechanisms to produce turbulence. The turbulence produced induces an upward buoyancy flux, raising the overall centre of gravity of the lake. Up to 90% of this mechanical energy is lost to dissipation.

The vertical model structure of basin scale waves, described above, appears to depend on the wind forcing and the basin slope, but most of the energy resides in the first and second modes. The presence of large basin scale gyres has been demonstrated by Ivey and Maxworthy (1992), the role of intrusions has been documented by Lemckert and Imberger (1993; 1995) and the importance of internal waves has been realized by Oldham and Imberger (1993). For a descriptive account of these motions and pertinent processes, the interested reader should peruse Imberger (1998, 1994), and for more details Imberger and Patterson (1990), Imboden and Wuest (1995), Gargett (1989) and Fernando (1991), articles on physical limnology, advection and mixing in lakes (and oceans), under neutral and stratified conditions, Ottino (1990), a review article on chaos, Hutter (1991), Gregg (1989) and Gregg *et al.* (1993), articles on the theory of internal seiches and waves and how the latter lead to mixing in a stratified environment, the Proceedings of three specialty workshops on internal waves held in Hawaii several years back, as well as Garret *et al.* (1993) and Lemckert and Imberger (1998), articles on the benthic boundary layer mixing.

1.2 Vertical structure of a lake

As documented by field work, laboratory work and numerical simulations, a lake can be partitioned in: a) a turbulent surface layer, b) a turbulent sub-surface layer, c) a quiescent lake body (main water column), and d) a partially turbulent benthic layer. The flow regime in these four regions (as stated above) depends greatly on the stratification, the present meteorological conditions and the amount of *ringing* in the lake (Imboden, 1990). In the classical limnological literature (Wetzel, 1975), the main water column is separated into two regions, the metalimnion and the hypolimnion. In that respect, the thermocline is defined as the surface where the buoyancy frequency N is a maximum, a surface embedded in the metalimnion; yet, such a precise separation is only artificial because of the gradual decay of frequency N with depth. Water in the metalimnion and hypolimnion is predominantly laminar but, periodically generated, localized bursts of turbulence transport vertically mass, momentum and energy.

The surface layer is defined as the layer of water extending from the free surface, where waves form and break, down to a depth where the direct influence of the surface wind and surface fluxes stop. Near the free surface, a shear layer exists which is stirred by the wind turbulence, wave breaking, and shear-produced turbulence. Turbulence in this (near surface) layer is exported to deeper water by turbulent diffusion, vertical jets associated with wave breaking, and coherent motions such as Langmuir and penetrative circulation. At the lower extreme of this import (or surface) layer a relatively sharp temperature gradient is formed, the diurnal thermocline, accompanied by an elevated velocity shear with associated shear production of turbulence; across this lower shear layer billows form, due to Kelvin-Helmholtz instabilities, and cause further mixing. The near surface and input layer constitute the diurnal surface layer, also known as either surface mixed-layer or surface mixing layer (SML), as often times this layer is not completely mixed but only mixing. In the remaining of this paper we may use both terms indiscriminately.

Under certain wind conditions, a second surface layer forms below the diurnal thermocline, 2-5 m thick, known as subsurface mixing layer (SSML), where the dissipation rate of turbulent kinetic energy, ε, remains nearly constant (Imberger, 1998a).

Separate, simple models of the integrated turbulent kinetic energy (TKE) balance, in the surface and billow layers, using several coefficients representing efficiencies of energy conversion, may be obtained by considering ***first*** the balance among the change of storage of TKE, heat convection and wind stirring inputs, the TKE transport and turbulence dissipation, and ***then*** the balance among the TKE import to the billow layer, the shear production at the base of the surface layer, the spin up of TKE, the entrained potential energy flux and the internal energy wave leakage. Partial balances between individual terms in these two equations lead to various deepening laws (of the surface layer) which express the surface layer depth as a function of time.

The surface and subsurface mixing layers are important for the growth of plankton. Since plankton grows in the surface layer in response to light, any coherent motions in the surface layer (which sweep plankton through a cyclic light regime) are certainly important.

The (turbulent) benthic boundary layer, at the bottom of stratified lakes, has thickness anywhere from a few centimeters to a few meters that maybe approximated as 0.15λ, where λ is the wave length of the incident internal wave measured perpendicular to the bottom slope (Ivey *et al.*, 1998). The dissipation level there varies between $10^{-8} \div 10^{-5}$ m^2s^{-3} and turbulence may originate from two sources. First, basin scale currents lead to a turbulent boundary layer, and second modes or internal waves impinging on the lake sloping boundary may break generating turbulence. The first mode basin scale seiches are, thus, damped by both boundary friction and non-linear steepening (Lemckert *et al.* 1998). Wave breaking releases TKE driving a flux, in the benthic boundary layer, which in turn cycles the hypolimnetic water through the latter (the benthic boundary layer, i.e.).

Although our understanding of transport processes in lakes (and in other aquatic environments) has been improved in the recent years, some uncertainty may still exist, particularly as regards the *flux path* of nutrients in the water body (see also Imberger, 1998a, and Papadimitrakis and Chioni, 2011), the role of rate of strain to which plankton is exposed, the turbulence induced by surface wave breaking and the description of coherent motions, that is uncertainty among the mechanisms which provide a detailed link between the physical and biological processes occurring in lakes {see also Imberger, 1998a; Imboden, 1998; and Papadimitrakis, 2005}.

1.3 Internal wave and other lake motions

Internal waves may arise from pressure fluctuations at the base of the surface layer (Wijesekera and Dillon, 1991; Zic and Imberger, 1992) or may originate either from the top of the benthic boundary layer (which mirrors the behavior of the surface layer) or from turbulent bursts within the fluid itself (Taylor, 1992). They are one of the most ubiquitous features of stratified flows and, although capable of propagating over long distances and through great depths, they tend to be concentrated in the regions of strong buoyancy forcing, i.e. at the base of the surface layer. Internal waves have much smaller frequencies, but much larger amplitudes, than those of surface waves and, among them, those with long wave periods travel almost horizontally and appear as intrusions. In large lakes, where rotation is important, long internal waves may induce large *pancake* shaped eddies which traverse the lake and carry mass with them. Thus, internal waves play a major role in lake dynamics as they are an energy source for the turbulent transport within the water body, the driving force of the buoyancy flux in the benthic boundary layer, and the energy source for the multitude intrusions and (synoptic) eddies found in the metalimnion and hypolimnion, which transport water horizontally over large distances. Energy moves, with internal waves in the form of *wave packets*, in the direction of wave group velocity.

Currents, eddies and internal waves combine to trigger patches of turbulence throughout the lake. Turbulent bursts, by transporting vertically mass, also lead to the formation of (horizontal) intrusions and eddies whose vertical vorticity, strengthen at the expense of horizontal vorticity, leads to an upscale growth of these eddies via the Biot Savart law (Narimousa *et al.*, 1991). Turbulent bursts constitute a mechanism which concentrates locally, at any particular time, the energy density (in hypolimnion) and causes a local breakdown and mixing. The vertical transport of mass by these bursts creates an imbalance in the density field, sets up a horizontal density gradient and induces a horizontal intrusive flow which, propagating into the fluid, carries mass with it. This explains how the bulk of horizontal dispersion takes place in deeper regions of lakes. Intrusions must negotiate a fluid, which itself is filled with other intrusions, eddies and internal waves that may also contain a background shear. A detailed description of all mechanisms contributing to the bursting sequence (a sequence of stages where most of turbulence is produced) in the vicinity of an air-water interface, roughened by wind-generated waves in the presence of swell, can be found in Papadimitrakis *et al.* (1988).

The picture obtained from power spectra of temperature (and velocity) fluctuation measurements in lakes suggests that energy from the surface waves is continuously fed into turbulence, in the surface shear layer, and subsequently spread in the mixed-layer. Internal wave-wave and wave-shear current interactions result in transferring energy to motions of decreasing scale made up of internal waves, eddies and intrusions. At a certain scale, the local shear associated with the motion becomes sufficiently strong to cause either a new shear or a convective instability, which then leads to turbulence. The turbulent motion absorbs some of this energy produced and transfers it to even smaller scales where it is finally dissipated. The eddies and intrusions, so generated, lead to strong horizontal mixing, and turbulence contributes mainly to vertical mixing.

1.4 Mixing mechanisms in lakes

The above picture confirms earlier observations (Imberger and Patterson, 1990) suggesting that the mixing mechanisms in lakes appear to fall into three major classes, such as mechanisms of internal wave breaking, mechanisms of Kelvin-Helmholtz billowing formation and mechanisms producing gravitational overturning. First, internal wave-wave interaction leads, under the right conditions, to a growth of one wave-length at the expense of the others until breaking occurs. These interactions may involve triads of internal waves, parametric instabilities or wave-wave-shear interaction. Second, the local shear may be raised, by the combining of long and short internal waves, to such a level that Kelvin-Helmholtz billowing can take place. Thirdly, gravitational overturning can be induced by absorption of wave energy at critical layers.

Turbulent mixing takes place at very fine scales. The effective flux resulting from a turbulent event may be parameterized by the local turbulent Froude, Fr_t, and Reynolds, Re_t, numbers, regardless of the age or origin of the particular turbulence.

For the biologist and chemist mixing and advection is more important. Periodic seiches and internal waves are certainly important, since they provide energy for mixing, but are in themselves of no great consequence for the biological or chemical systems since they do not distribute mass in general.

In summary, the processes which transport water masses (and also mix pollutants), nutrients and other living organisms in lakes include: wind-induced, convective and billow deepening of the surface and subsurface mixed-layers, upwelling, differential deepening, differential heating and cooling in space and/or time, insertion, entrainment, diffusion, gravitational overturning of internal waves and plunging of stream inflows in the hypolimnion. Due to the limited space (allowed), we have not expanded on upwelling and some other processes (as insertion, entrainment, diffusion and the plunging of inflows,...).

In the following, the time and length scales of basic physical processes occurring in lakes are briefly presented in an ecohydrodynamic perspective, identifying appropriate spectral windows and the non-dimensional parameters governing the various flow regimes, and emphasizing their relative importance in lake ecosystems. Spectral window (and sub-window) identification follows in section 2. In section 3 we present the turbulent length and time scales associated with the basic physical processes, and the relevant non-dimensional parameters which determine the flow regime and govern the hydrodynamics and thermodynamics in lakes. For completeness, we present, in section 4, some aspects of ecological time and length scales. A summary of the most important spatio-temporal scales and dimensionless parameters governing the lake dynamics is presented in section 5.

2. SPECTRAL WINDOWS AND HYDRODYNAMIC VARIABILITY

Although, due to the non-linearity of the governing hydrodynamic equations, a great number of length and time scales of motion exists in lakes (and in other aquatic environments), certain domains of these scales dominate the dynamics of motion of the aquatic systems. These *spectral windows* correspond to the scales of external forcing (energy inputs) or of intrinsic mechanisms (eigenmodes - Papadimitrakis and Nihoul, 1997a). The basic hydrodynamic equations contain four characteristic eigen frequencies:

I. The viscous cutoff frequency, f_V (or inversely the cutoff wave number, k_V), which is a measure of the effects of molecular diffusion and viscous dissipation of TKE; $f_V \sim (\varepsilon/v)^{1/2}$ and $k_V \sim (\varepsilon/v^3)^{1/4}$ (where v is the kinematic viscosity of water), with typical values of f_v^{-1} of about (1 ÷ 10) s and k_v^{-1} of about ($10^{-5} \div 3\text{x}10^{-5}$) m. Components of motion with time scales smaller than f_v^{-1} and length scales smaller than k_v^{-1} are dominated by molecular diffusion and dissipation processes.

II. The buoyancy (or Brunt-Vaisala) frequency, N, which is a measure of the stratification. $N^2 = -(g/\rho_m)d\rho_e(z)/dz$, with typical values of N^{-1} of about ($10^2 \div 10^4$) s; the minus sign has been incorporated in the above expression of N^2 to make the latter positive when stratification is stable. Components of motion with time scales of $O(N^{-1})$ are affected by stratification. Here, g is the gravitational acceleration and ρ_m is a characteristic mean water density, in the vertical direction, that corresponds to the mean water temperature. Typically, the water density maybe represented by the sum of three components, namely: $\rho = \rho_m + \rho_e(z) + \rho'(x, y, t)$, where $\rho_e(z)$ corresponds to the density structure (above ρ_m) caused by temperature variations alone in the vertical direction (and in the absence of motion), whereas the fluctuations ρ' in the horizontal plane (above the sum of the other two density components) are induced by any motion. In the presence of thermal stratification with a sharp density (or temperature) gradient in the metalimnion region, the density gradient $d\rho/dz$ maybe approximated as $\Delta\rho/\Delta z$, where $\Delta\rho$ represents the density difference in the base of pycnocline/thermocline that spans over a vertical distance Δz. Imboden (1998) has provided alternative (but more complicated) expressions for N^2 which account for the presence of various solutes in the water and the isentropic heat transport across the layer (see also Millard *et al.*, 1990).

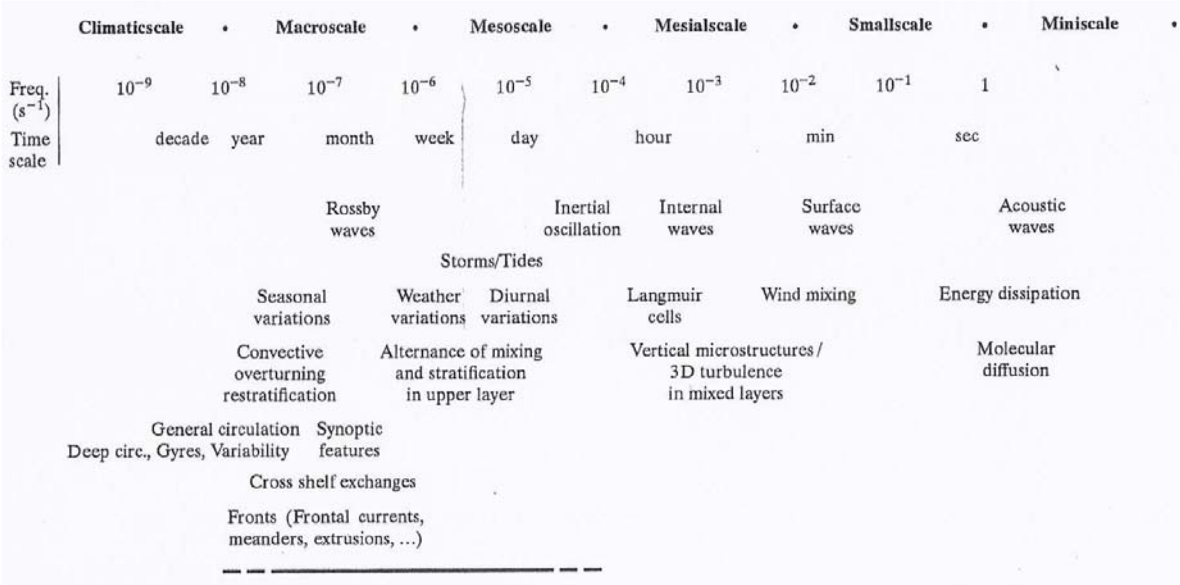
III. The Coriolis frequency, f_c, which is a measure of the influence of the earth's rotation on water motion; f_c is defined as twice the vertical component of the earth's rotation vector; in mean latitude, $f_c \sim 10^{-4}\ s^{-1}$. Components of motion with time scales comparable to or greater than f_c^{-1} (i.e., a few hours) are affected by the earth's rotation.

IV. The Kibel frequency, f_K $(= \beta r)$, a measure of the influence of the earth's curvature on water motion; β is the gradient of f_c, r $(= NH/f_c)$ is the Rossby radius of deformation and H is a typical water depth. The typical/maximum value of f_K is on the order of $10^{-6}\ s^{-1}$. Components of motion with time scales comparable to or larger than f_κ^{-1} (a few weeks) are affected by the earth's curvature.

External forcing at the air-water interface is characterized by diurnal, weekly (time scale of the wind field's variability) and seasonal variations of momentum, heat and mass fluxes, with typical frequencies of $10^{-4}\ s^{-1}$, $10^{-6}\ s^{-1}$ and $10^{-7}\ s^{-1}$, respectively. A schematic representation of aquatic hydrodynamic variability, as a function of characteristic frequencies or time scales, is given in Table 1 (see also Papadimitrakis and Nihoul, 1997b). In general, time and length scales are related and it is customary to associate high frequencies and high wave numbers, although the association may be different for eigenmodes and forced oscillations.

The transfer of energy among spectral windows is affected by non-linear interactions. The latter are present -with varying efficiency- at all scales. However, it appears that there is no *cogent* volume force acting in the small scale range of frequencies $10^{-2}\ s^{-1} < f < f_k \sim 1\ s^{-1}$. Thus, except for surface waves and other processes occurring at the air-water interface, the spectral window $10^{-2}\ s^{-1} < f < 1\ s^{-1}$ is dominated by non-linear energy transfers from one scale to another producing -with the multiplication of interacting eddies of various scales- a state of macroscopic multi-dimensional chaos, i.e. turbulence. In the mesoscale (mesial scale) range of $10^{-4}\ s^{-1} < f < 10^{-2}\ s^{-1}$, turbulence is still random.

Table 1. Schematic Representation of Aquatic Hydrodynamic Variability



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At larger scales, restoring mechanisms oppose the tendency of non-linear interactions to create disorder and turbulence is inhibited (as stratification flattens out eddies into *blinies* and reduces vertical mixing) and less disorganized (since vertical overturning is imposed on synoptic scale *rossbies*). A *bliny* (from the Russian *blini*) is a pancake-shaped eddy contributing to an energy cascade to smaller scales via epidermic instabilities and internal waves. A *rossby* (from the scientist Rossby) is a pseudo two-dimensional eddy column with scales of the order of the Rossby radius of deformation.

Turbulent features, like *blinies* and *rossbies,* are closely related to the eigenmodes of oscillations which can propagate in the corresponding range of scales (of internal waves, inertial oscillations, and rossby waves) as *wave packets*, resulting from non-linear interactions of such oscillations. The existence of eigenmodes of oscillations, in specific frequency ranges, also reveal the presence of well-defined restoring forces (associated with the vertical stratification, the earth's rotation, the earth's curvature,..and) opposing the tendency to create hydrodynamic disorder (turbulence of all scales), characteristic of non-linear, scale-cascading interactions.

The competition of structuring and destructuring mechanisms results in the cohabitation, in the aquatic system, of zones of significant (vertical or/and horizontal) gradients and well-mixed regions (turbulent, surface- and benthic-boundary layers, intrusions, intermittent and irregular blinies, rossbies) closely interrelated via stability/instability mechanisms. Semi-persistent physical structures, associated with vertical stratification (pycnoclines, microstructures, layers, boundary surfaces,...) and the horizontal frontal boundaries between adjacent water masses (upwelling fronts,...), are of great importance for the dynamics and structures of biological populations.

The temporal and spatial scales associated with the whole range of motions in lakes vary from seconds to months and from millimeters to kilometers, respectively. Currents and eddies vary slowly and are aperiodic, internal waves are periodic and have frequencies from N (a few min^{-1}) to f_C (a few $days^{-1}$), and turbulence is made up of motions (where buoyancy does not influence the motion) with frequencies from $(\varepsilon/v)^{1/2}$ (a few sec^{-1}) to N (a few min^{-1}). Thus, it becomes apparent that the time scales of importance in lakes cover several spectral windows and correspond to processes ranging from small-scale to macro-scale.

3. TURBULENT LENGTH AND TIME SCALES IN LAKES

3.1 Governing non-dimensional parameters

In turbulent flows, it is perhaps possible to identify some fundamental variables in terms of which all other characteristics of the flow may be described. Such variables indeed exist and are as follows:

S (s^{-1}), N (s^{-1}), ν ($m^2 s^{-1}$), κ ($m^2 s^{-1}$), t (s), q_i (ms^{-1}), l_{ci} (m)

where S is the horizontal velocity shear of the flow (or Prandle frequency), κ is the molecular diffusion coefficient, t is the time, q_i is the initial rms of the velocity fluctuations and l_{ci} is the initial rms scale of the motion (see also Papadimitrakis and Imberger, 1996). Five dimensionless parameters, which determine the flow regime, may be formed from these seven variables and are given as:

$Ri = N^2/S^2$; $T = S.t$; $Pr = \nu/\kappa$; $Fr_{ti} = q/(Nl_{ci})$; $Re_{ti} = ql_{ci}/\nu$

where Ri, Pr, Fr_{ti}, Re_{ti}, and T are, respectively, the (gradient) Richardson number, the Prandtl number, the initial turbulent Froude number, the initial turbulent Reynolds number, and a dimensionless time. The initial turbulent Froude and Reynolds numbers express the influence of the initial conditions on the evolution of the flow field. Another important dimensionless parameter is the flux Richardson number, R_f, defined as either –G/P, the minus ratio of buoyancy production of T.K.E., G, and shear production, P, or as –G/(G+P). It may also be defined as the product of Ri and the ratio of buoyancy diffusivity (a sub-window scale of vertical diffusivity of buoyancy) and momentum diffusivity (a sub-window scale of vertical diffusivity of momentum). R_f is also a measure of mixing efficiency in an aquatic environment (Imberger, 1998a).

Based on the above fundamental variables, the following two length scales maybe derived in terms of ν and N, namely: l_c (= g'/N^2) and $l_p = (\nu/N)^{1/2}$. The first scale represents the ensemble average of the largest eddies; it is termed the overturn (or reordered or displacement, or integral) scale and captures the largest observed motion. The second is called the primitive scale; g' (= $g\Delta\rho/\rho_o$) represents an effective reduced gravity across the base of the surface layer, Δρ is the density anomaly, i.e. the density jump across the base of that layer, and ρ_o is a reference (say the hypolimnion) density.

The dissipation of TKE ε, at a point, is a further parameter of the flow field. Using dimensional analysis, it is possible to express both l_c and ε in the form:

$l_c = (\nu/S)^{1/2} f_1(Ri, T, Pr, Fr_{ti}, Re_{ti})$; $\varepsilon = \nu S^2 f_2(Ri, Pr, T)$, (for $Ri \leq 0.21$)

Four more length scales may now be derived in terms of ε, κ, and ν. These are the Ozmidov, the Grasshof, the Batchelor and the Kolmogorov scale, l_o, l_g, η_B and η_V, respectively; namely:

$l_o = (\varepsilon/N^3)^{1/2}$; $l_g = (\nu^2/g')^{1/3}$; $\eta_B = (\nu\kappa^2/\varepsilon)^{1/4}$; $\eta_V = (\nu^3/\varepsilon)^{1/4}$

The Ozmidov scale is the vertical length scale (distance, i.e.) at which the buoyancy forces are of the same order of magnitude as the inertial forces. It is the largest possible *active* turbulent scale. The Grasshof scale represents the distance at which l_p is equal to l_c. The Batchelor scale is the smallest of all, and is the scale at which the density gradients are annihilated. Further vertical scales can be derived for the dissipation rate of scalar variance (Gibson, 1991); l_o, η_V and η_B can also be expressed as:

$l_o = l_p f_2^{1/2}$; $\eta_V = l_p Ri^{1/4} f_2^{-1/4}$; $\eta_B = l_p Pr^{-1/2} Ri^{1/4} f_2^{-1/4}$

One final length scale is the Ellison scale defined as: $l_E = -\left(\overline{\rho'^2}\right)^{1/2} / \partial\rho_m/\partial z$. Yet, Itsweire *et al.* (1993) has demonstrated that, with the exception of flows with very high mean flow Richardson numbers, where turbulence is not active, l_c and l_E can be taken as the same scale for practical purposes.

The pertinent time scales associated with the list of fundamental variables are:

$T_f = N^{-1} = (l_c/g')^{1/2}$; $T_m = g' l_c/\varepsilon$; $T_d = l_c^2/\nu$; $T_a = (l_c^2/\varepsilon)^{1/3}$; $T_V = (\nu/\varepsilon)^{1/2}$

Here T_f is the falling (particle) time scale, the time taken for particles, having a gravitational acceleration anomaly, to return to their stable equilibrium position by falling under gravity a distance l_c; T_m is the mixing time scale, the time it takes for the density anomaly to be mixed by the smaller scale turbulent motions; T_d is the molecular diffusion time scale; T_a is the advection time scale (if turbulence is active), and T_v is the Kolmogorov (or viscous) time scale. T_v also represents the inverse rate of strain γ $\{=(\varepsilon/\nu)^{1/2}\}$ of the smallest scale turbulent motion.

The hierarchy of these length scales (l_C, l_O, l_p, η_V, η_B) is very important in determining both the turbulent structure, and in affecting the various biological processes occuring in a lake. If $l_O > l_C$, there is more energy being dissipated than is implied by an energy cascade which commences at the largest scale where $Fr_t = 1$. The turbulence can, thus, grow with time until l_C reaches l_O. On the other hand, if $l_O < l_C$ dissipation active in small scales is smaller than would be expected from the energy cascade using l_C as the largest scale; active (inertially influenced) turbulence cannot be occurring at all of the length scales in the flow field and the motion can only be turbulent out to scales where $Fr_t = 1$. Scales larger than l_O represent motion with statistically unstable density fields, which may or may not be convectively unstable. Further, if the scale l_C approaches η_V, then the motion is completely damped by viscosity and buoyancy.

The ratio of the above length and time scales can be used to define a series of very important dimensionless numbers, namely:

$$l_o/l_c = Fr_t^{3/2};\ \ l_c/l_p = Gr_t^{1/4};\ \ l_c/l_g = Gr_t^{1/3};\ \ l_o/\eta_V = Fr_\gamma^{3/2};\ \ l_c/\eta_V = Re_t^{3/4};\ \ \eta_V/\eta_B = Pr^{1/2}$$

$$T_f/T_m = Fr_t^3;\ \ T_d/T_a = Re_t;\ \ T_d/T_f = Gr_t^{1/2};\ \ T_f/T_V = Fr_\gamma$$

where: $$Fr_t = \left(\frac{\varepsilon}{N^3 l_c^2}\right)^{1/3};\ \ Re_t = \left(\frac{\varepsilon\, l_c^4}{\nu^3}\right)^{1/3};\ \ Gr_t = \frac{N^2 l_c^4}{\nu^2};\ \ Fr_\gamma = \left(\frac{\varepsilon\, l_c}{g'\,\nu}\right)^{1/2}$$

and (for active turbulence) the basic definitions of turbulent Froude, Reynolds and Grasshof numbers, Fr_t, Re_t, and Gr_t have been used respectively. Fr_γ is the strain Froude number; it is the ratio of the rate of strain, γ, as measured by the dissipation at the smallest scales, and the rate $1/T_f$ at which gravity returns the unstable density anomalies to their original positions. Since this latter rate is the maximum rate at which internal waves can move the fluid, the value of Fr_γ indicates whether the rate of strain is influenced by internal wave activity. These non-dimensional parameters may be given different physical interpretation depending on the expression used.

If $Fr_t > 1$, turbulence is active in the sense that inertia, measured by the dissipation, is large enough at all scales to support motions in the stratified background; entrainment can occur leading to even larger scale motions. The available potential energy (from the density anomalies) does not add much to the energy budget and the density anomalies disappear, not by the fluid returning to its neutral position but rather by turbulent mixing.

If $Fr_t < 1$, the converse is true. The kinetic energy of the motions, as measured by the dissipation, is smaller than that present in the large scale convective unstable density perturbations; energetic turbulence is confined to scales which are smaller than l_O, which itself is smaller than l_C. The density anomalies are not erased by mixing before the fluid can return to its equilibrium position and we may expect a mixture of large scale gravitational oscillations, convectively driven thermals and small scale active turbulence sustained by both residual inertia and by convection. Now the history of the flow is important and all variables depend on the initial length scale, l_{Ci}, and the initial dissipation rate ε_i. Then, $Fr_{ti} = 1$ and $l_{Ci} = [\varepsilon_i/N^3]^{1/2}$ (Imberger and Ivey, 1991; Ivey and Imberger, 1991).

If Fr_γ is large, the fluid is straining much faster than can be accounted for by internal wave activity. The flow field evolves too rapidly for internal wave activity to smooth out the density anomalies so that active turbulence must be present to explain the rapid rate of strain. If Fr_γ is small, internal waves can adjust the density anomalies back to their original position. The importance of this observation lies in the fact that turbulence causes mixing, whereas internal waves can only transport momentum, not mass. Since Fr_γ determines the ratio l_c/η_v, it is clear that the value of Fr_γ is also a measure of the scale separation between the smallest and the theoretically largest scale.

Given the above definitions of Fr_t, Re_t, Fr_γ and Gr_t it is possible to construct an Fr_t - Re_t turbulence activity diagram for the measured dissipation, ε, and density anomaly, Δρ, with parametric dependencies on Fr_γ and Gr_t. This is shown in Figure 1 (see also Papadimitrakis and Nihoul, 1997a); the type of motion is reflected by the position of the data point on this diagram. The flux Richardson number, R_f, appearing in the same diagram, is defined as the minus ratio of buoyancy and mechanical energy production of turbulence (i.e., -G/P).

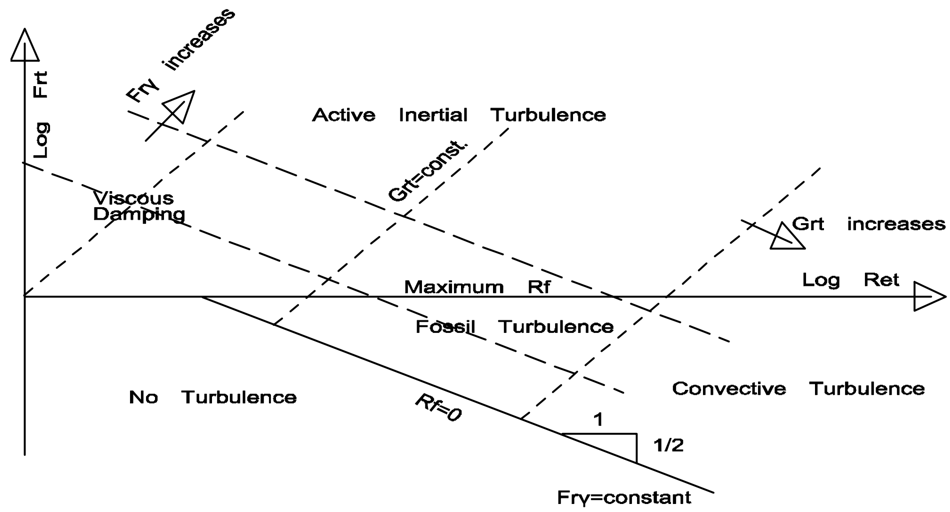

Figure 1. Activity diagram for turbulent shear flow of a stratified fluid

Ivey *et al.* (1998) have provided detailed analytical expressions relating either R_f, Fr_t and Re_t or R_f, Fr_γ and Re_t, or R_f and Fr_t. These authors found that $R_f = 0$, when $Fr_\gamma = 15^{1/2}$, and in the upper limit, when $Re_t \geq 75$ and $Fr_t = 1$, $R_f = 0.28$. The arguments above do imply that a number of regimes, in the Fr_t - Re_t parameter space, are possible. Ivey *et al.* (1998) identified indeed five such regimes (for $Pr > 1$, as shown in their Figure 5 and their Table 1), similar to those shown in our Figure 1. One of the advantages of Figure 1, and the explicit expressions provided by Ivey *et al.* (1998), is that they allow estimation of the mixing efficiency, R_f, down to and including the limit where the combined action of buoyancy and viscosity suppress the buoyancy fluxes, as observed in many lakes and in the ocean.

Lemckert and Imberger (1998) have also provided a more simple (Fr_t - Re_t) activity diagram showing the nature of turbulent (mixing) events within turbulent benthic boundary layers generated by shoaling internal waves, a structure displaying the transitions from the inertia dominated to buoyancy dominated turbulence and to internal wave motions. Fr_t and Re_t numbers were found, in their study, to vary from 5×10^{-1} to 10^2 and from 10^{-1} to 5×10^3, respectively. Direct numerical simulations of stratified turbulence, performed by Yamazaki and Ramdsen (1998) for comparing dynamic conditions of laboratory and field observations, have shown that Fr_t and Re_t numbers vary from 5×10^{-2} to 10^1 and from 10^{-1} to 10^5, respectively, with laboratory observations to remain dynamically similar to counterpart field observations when the non-dimensional parameter ε/vN^2 $(= T_f/T_v)^2$ remains less than 160.

Lake activity can also be classified with the aid of Froude, Wedderburn and Lake numbers. The objective, underlying these classification schemes is to provide the biologist and chemist with a simple single parameter which summarizes the activity or strength of a particular dynamical process. The inflow and outflow Froude numbers, Fr_I and Fr_O, classify the inflow and outflow dynamics, the Wedderburn number, W, the rate of deepening and tilting of the bottom interface of the surface layer, and the magnitude of the Lake number, L_N, is an indication of the mixing activity in the deeper part of the lake. These numbers have been statistically correlated with biological and chemical variables in lakes. L_N, for example, is correlated well with the concentrations of oxygen, manganese and iron in the hypolimnion (Robertson and Imberger, 1994), W with algal growth patterns and Fr_I

with algal species. All of the above numbers also serve to put into context the relative strength of competing influences. Furthermore, it has been argued that the severity of large basin scale Kelvin, Poincare and/or forced gravity waves is inversely proportional to the magnitude of the Wedderburn, W, and Lake numbers L_N (Imberger, 1998a).

L_N and W, Fr_I and Fr_o are defined as follows:

$$L_N = \frac{(z_g - z_0) M g \left(1 - \frac{z_T}{H}\right)}{\rho u_*^2 A^{3/2} \left(1 - \frac{z_g}{H}\right)} \; ; \quad W = \frac{g' h^2}{u_*^2 L'} \; ; \quad Fr_I = \frac{Q_I}{A\left(g^* H\right)^{1/2}} \; ; \quad Fr_o = \frac{Q_O}{\left(\frac{\Delta\rho}{\rho_0}\right)^{1/2} H^{5/2}}$$

where A is the lake surface area, M is the lake water mass, z_O and z_g are the heights to the centers of mass and volume, respectively, z_T is the height to the center of the metalimnion (all heights being measured from the lake's bottom) and H is the lake depth; h is the thickness of the surface layer, u_* the water friction velocity, due to wind-stress, and L' the fetch length; u_* maybe estimated as: $u_* = (C_d\rho_\alpha/\rho_w)^{1/2}U_{10}$, where C_d is the wind-stress coefficient, ρ_α and ρ_w are the air and water densities at the air-water interface, and U_{10} is the wind speed measured at the height of 10 m above the mean water elevation. With typical values of $C_d = 1.3\text{x}10^{-3}$ and $\rho_\alpha/\rho_w = 10^{-3}$, u_* maybe approximated as $1.14\text{x}10^{-3}U_{10}$; Q_I represents the inflow discharge and g^* is a reduced gravity based on the density difference between inflow and the surface water; Q_O is the outflow discharge and $\Delta\rho$, in the definition of Fr_o, represents the density difference between the surface and the outflow level, whereas ρ_o is the water density at the outflow level.

A similar inflow lake number, L_{NI} may be defined in terms of Q_I, z_0, z_g, z_T, A and the density of the inflow. Finally, an underflow Froude number (also called internal Froude number) may be defined, as Fr_I, in terms of the underflow discharge velocity, the density difference between the lake and the underflowing water, and the hydraulic depth of the underflow.

4. ECOLOGICAL TIME AND LENGTH SCALES

4.1 Ecosystem windows

In aquatic ecosystems, geo-chemical and biological (i.e. ecological) processes can also be characterized by appropriate time and length scales. Bio-ecological systems can be studied at different length scales, from individual organisms to a whole population. Various studies (Steele, 1978; and others) show time scales characterizing individual organisms from days to years, associated with length scales ranging from a few microns to less than a meter. The time scales of a whole population aggregate are of the same order of magnitude as the time scales of the individuals forming that population (if they are not too different), but the associated length scales are defined by the sizes of habitats and patches, and the associated variances and gradients.

Fortunately, in these ecosystems a hierarchical organization exists which results from the different rates of ecological processes occurring in the multi-scale physical environment. Processes with similar time scales belong to the same levels of hierarchy. Phenomena at a particular level are, to a large extent, dissociated from lower level ***noise*** or higher level *global trends* and may be relatively easily singled out of the total complexity of the ecosystem (O'Neill, 1989). Thus, ecological processes can be analyzed as comparatively simple systems, when viewed through an appropriate range of time and length scales (a spectral window) of the ecosystem.

Lakes and other aquatic ecosystems have *endogenous* time scales which determine the hydrodynamic processes that may significantly interact (in the framework of several biochemical processes) with populations of the aquatic communities. Hydrodynamics maintain a permanent strain on the ecosystem which (the strain, that is) tends to impose to the latter the length scales of the synchronous physical mechanisms (Denman and Powell, 1984; Nihoul and Djenidi, 1990). Thus, important links and interactions between hydrodynamics and bio-geochemical processes do exist in the aquatic environment.

Since most of the practical problems encountered in aquatic ecosystems arise at the level of the ecosystem, attention ought to be focused on those time scales which are inherent to the ecosystem. This usually implies periods of time varying from 10^4 s - 10^8 s (or 10^5 s - 10^7 s), which correspond to characteristic cycles in the life of many pelagic and benthic populations (periods of diurnal, seasonal and annual oscillations) and to the rythm of human activities, interplaying, for better or worse, with the aquatic system.

The corresponding hydrodynamic processes in the above range of scales, namely the mesoscale (with a typical time scale t_C of a few hours), the synoptic scale (with t_C = a few days) and the seasonal scale (with t_C = a few weeks) processes, which constitute the *weather of the aquatic environment*, are likely to interact with the various ecological phenomena. Consequently, motions at the time scale of the *weather of the aquatic environment* are expected to resonate with the ecosystem dynamics, and (as cited above) to maintain a permanent strain on it through the advection process. Stated differently, the time scales which characterize the spectral windows of an ecosystem are function of the ecosystem's behavioral rates, the associated length scales of which are set by the resonant hydrodynamic forcing through the process of *ecohydrodynamic adjustment* (Nihoul and Djenidi, 1991).

Aside from the time and length scales -of the hydrodynamic processes associated with external or internal forcing mechanisms occurring in lakes (and other aquatic systems)- cited in Table 1 and described previously, the interaction processes (between hydro- and eco-dynamics) can also be characterized by specific time and length scales. The comparison of scales between these (family of) processes indicates which processes are actually in competition in the aquatic environment. At hydrodynamic time scales much smaller than the interaction scales, very little interaction takes place over times of significant hydrodynamic changes, and the various constituents are essentially transported and dispersed *passively* by the water. On the other hand, hydrodynamic processes with time scales much larger than the interaction time scales scarcely affect the dynamics of interactions over any time of interest. Only those processes which have time scales comparable with the interaction time scales can significantly affect the bio-geochemical interactions and act to constraint the activity of chemical and biological systems.

Thus, any particular biochemical process must be studied in the framework of its *spectral window*, subject to the *resonant* hydrodynamic constraint, embedded in the slowly varying environment of the larger scales and blurred by the (non-linear) diffusing effect of *sub-window* or *sub-grid* scale-turbulent or *pseudo-turbulent* fluctuations. Once the time scales of the ecological processes of interest are identified, the appropriate *spectral window* is determined. The hydrodynamic processes which are responsible for the transport and space-time distribution of the ecological state variables are the ones which have the same time scales with the ecological processes. The length scales of these *resonant* hydrodynamic processes are imparted to the ecosystem by the persisting non-linear constraint of their embedment in the flow field. The spectral window of plankton activities, for example, includes time scales in the band ($10^4 \div 10^7$) s, even though the individual plankton organisms have time and length scales in the viscous regime (s, mm).

Sub-window hydrodynamic scale processes include the viscous dissipation regime, surface waves, eddy and bliny turbulence, Langmuir cells, internal waves, and inertial oscillations. Thus, miniscale, smallscale, mesialscale and the faster mesoscale processes provide the physical background for the main ecological interactions, at population scales, and have a direct effect only on the dynamics of the lower levels of the ecological hierarchy. The influence of lower level physical and biological processes on the large- *window scale* - components (the residue of non-linear interactions) can be parameterized in terms of simple concepts, such as diffusion and mixing, resource supplies, etc., that result in fluxes (in physical space) and translocations (in state space) of matter and energy. It must be emphasized that the accumulation of all of the *sub-window* scale fluctuations results in, relatively, smooth residual forcing on higher hierarchical levels.

4.2 Other scaling concepts

During the last decade or so several scientists, mainly, from the 'Biology' community have also promoted concepts related either to length (and to a lesser extent to time) scales or to non-dimensional parameters that differ from the ones described previously, in some detail. Yet, when these concepts are viewed more scholastically it becomes evident that, to some extent, they are related to the scales and parameters introduced before. Hence, it was felt appropriate to introduce them here, briefly, without any further exploitation or other reference.

Keeling *et al.* (1997), for example, utilizing 'fluctuation analysis' identified characteristic length scales (in spatial models of ecology) which provide an optimal length size for extracting non-trivial large scale behavior in spatial biological models, one of which refers to a complex marine system. The latter model is a cellular automation that covers a complex statistical structure. Cellular automata (c.a.) are discrete dynamical systems in which many simple components act together, locally, to produce complex patterns on a global scale, which may exhibit 'self-organizing' behavior (Mynnet and Chen, 2004). Owing to the ability of c.a. to model local interaction and spatial heterogeneity, they have been applied to very broad fields, as for example in a confined ecosystem or in open aquatic ecosystems where external forcings are accounted for in the definition of cell state transitions. Results show that c.a. could be a valuable paradigm in ecological and ecohydraulics modeling.

Keeling *et al.* (1997) also showed: a) that 'fluctuation analysis' may also be used in the identification of aggregation or dispersal (of species) at various scales, b) that this method is rigorously justified when the system satisfies the Fortuin-Kasteleyn-Ginibre property and has a fast decay of correlations, and c) that the length scale 'fluctuation analysis' is related to the hydrodynamic limit method(s) for deriving large scale equations in ecological models.

Kemp *et al.* (2000) established the link between 'functional habitats' (i.e., biologically defined habitat units) and 'flow biotops' (i.e., hydraulically defined habitat units) using a properly defined Froude number. The approach was to examine the relationship between 'functional habitat' occurrence and Froude number, with the aim of developing 'habitat preference curves'.

Melbourne and Chesson (2005) implemented a systematic approach to the problem of 'scaling up', in the field of ecology, using scale transition theory; 'scaling up' from local scale interactions to regional scale dynamics is a critical issue. These authors showed that the dynamics of larger spatial scales differ from predictions based on the local dynamics alone, because of the interaction between local scale non-linear dynamics and the spatial variation of density in the environment. They also delineated the four steps of 'scaling up' process.

Woodson *et al.* (2005) studied the response of various species (as Copepods, etc.) to spatial gradients of flow velocity and fluid density in order to determine whether the presence of physical gradients initiated local search for resources or promoted aggregation. Experimental evidence showed a significant increase in the proportion of time (spent) in the gradient layer region, relative to the total time (spent) in the observation window (which is proportional to the residence time), in response to the velocity gradient layer. Behavioral changes, such as increased swimming speed and turn frequency, were consistent with area-restricted search behavior. Density gradients were found to act as barriers to vertical movement and not as a positive cue for, area restricted, search behavior. Thus, velocity and density gradients were found to play important, yet different, roles in defining patterns at fine-to-intermediate scales in zooplankton ecology.

4.3 Plankton time scales

Surface waves and breaking waves of large scale in particular, as well as Langmuir circulations, contribute to the vertical transport of plankton in a succession of downwelling-upwelling (circulation) cells. Since Langmuir cells rarely penetrate deeper than some dominant wave lengths (of about $\leq 10\ m$) with typical down and up-welling speeds on the order of a few cm sec^{-1}, it may be concluded that the characteristic time of plankton circling around is on the order of 10^3 s. With a time scale, t_C, and a length scale for the large energy containing eddies, l_C, based on reasonable estimates of TKE, E ($\approx 10^{-4}$ m^{2}s^{-2}), and of the dissipation rate, ε ($\approx 10^{-7}$m^{2}s^{-3}), on the order of (t_C = E/ε =) 10^3 s and ($l_C \approx E^{3/2}/\varepsilon$ =) 10 m, it may be also concluded that large scale turbulence circulates plankton in about half an hour. During periods of stratification, and with a reasonable buoyancy frequency of O(10^{-3}) s^{-1}, one can argue that, in the stratified upper layer, internal waves take over the up- and downwards displacement of plankton with much the same time and length scales as turbulence (Nihoul and Djenidi, 1990). With an order of magnitude lower estimates of E and ε, l_C becomes of O(1) but t_C remains essentially the same.

It must be emphasized that although the mixed-layer dynamics has a *sub-window* scale with $t_C \approx 10^3$ s, the succession of mixed-layer events: stratification, formation of the diurnal thermocline, erosion of the latter by wind mixing, deepening of the mixed layer, etc., has larger time scales pertaining to day-night variations, atmospheric weather changes... with $t_C \sim 10^5$ s, and is likely to resonate with certain aspects of ecosystem dynamics.

It is also understood that depending on whether the time scale, i.e. the life time of particular hydrodynamic processes (which perhaps cause mixing in a lake or other aquatic environment), falls in the *aquatic weather sub-window* scale range (or not), their effect (i.e., of such processes) on the ecosystem dynamics may be beneficial or catastrophic. Such comparison may also be extended to include relevant ecosystem length scales, as for example is the scale associated with the *critical depth*, where the depth integrated net productivity is zero. If the surface mixed-layer depth is greater than this *critical depth*, net productivity may be very low; otherwise it may be quite high (if nutrients are available).

5. CONCLUDING REMARKS

Two classes of temporal and spatial scales appear to be important in all lakes and other similar aquatic environments. The first class expresses the hydrodynamics and thermodynamics of the water enclosure examined and is related to the external and internal forcing mechanisms that occur there and form the basic physical processes. The other class is related to the biology-ecology of lakes, etc., and expresses the remaining bio-geo-chemical processes that also contribute either by themselves or in interaction (with themselves and with the lake hydrodynamics) to the shaping of the pertinent ecosystem. Our complementary attention focuses (more or less) on the ecological time and space scales, formed mainly by the biology of the lake, etc., and to a lesser extent by other active geo-chemical processes.

Thus, the basic physical processes occurring in lakes were first reviewed and the relevant length and time scales were then presented in an ecohydrodynamic perspective. Such length scales are: the overturn scale, l_C, the primitive scale, l_p, the Ozmidov scale, l_o, the Grasshof scale, l_g, the Kolmogorov scale, η_K, and the Batchelor scale, η_B. The hierarchy of these length scales is very important in determining both the turbulence structure, and in affecting the biological processes occurring in lakes. The pertinent time scales are: the falling (particle) time, T_f, the mixing time, T_m, the molecular diffusion time, T_d, the advection time, T_a, and the Kolmogorov (or viscous) time, T_v. For large lakes, two more time scales can be formulated, based on the Coriolis frequency, f_C, and the Kibel frequency, f_κ .

All of these scales (but f_C and f_κ) are defined in terms of the shear rate, S, the buoyancy frequency, N, the molecular viscosity, v, and diffusivity, κ, the initial r.m.s of the velocity fluctuations, q_i', the initial rms length scale of the motion, l_{ci}', and the rate of turbulent kinetic energy dissipation, ε. Combinations of the above parameters yield important dimensionless numbers, as R_f, Ri, Pr, T, Re_t, Fr_t, Gr_t, and Pr_t, the flux and gradient Richardson numbers, the Prandtl number, a dimensionless time, and the turbulent Reynolds, Froude, Grasshof and Prandtl numbers, respectively. A few more dimensionless numbers maybe also useful in characterizing the general dynamic behavior of lakes, namely: the Lake number, L_N, the Wedderburn number, W, the inflow lake number, L_{Ni}, the inflow and outflow Froude numbers, Fr_I and Fr_O, and the ratio, R, of Rossby radius of deformation to a lake's characteristic dimension. Ratios of the length and time scales provide also similar dimensionless numbers.

Respective scaling concepts relating to the ecology of lakes were also presented, as well as the time and space scales of plankton communities in comparison with the pertinent hydrodynamic scales.

The importance of the above dimensionless numbers in establishing the dynamic flow regimes in lakes, having different morphological characteristics and subject to various external forcing conditions, is examined in a companion paper (Papadimitrakis and Chioni, 2011). There, the scaling concepts elaborated here will be used in two lakes, located in the Northern Greece, and the response of these water bodies to the respective external (and/or internal) forcing will be examined in terms of the determined scales and dimensionless numbers.

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