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Interaction Between Hydrodynamics and Mass-Transfer at the Sediment-Water Interface

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Abstract: Modeling mass-transfer across the sediment-water interface is a significant issue in environmental hydraulics. In fact diffusional exchanges of solutes between the bed sediment and the overlying water column could greatly affect water quality. Particularly, diffusional flux of dissolved oxygen (DO) towards the bed sediments from the water column could be responsible for low and unacceptable levels of DO in the ecosystem. This flux depends both on sediment and flow characteristics. The objective of the present paper is to investigate the interaction between flow hydrodynamics and dimensionless fluxes of dissolved substances across the sediment-water interface. Therefore, some literature predictive models are compared with experimental laboratory data collected both in flumes and benthic chambers. Also, the influence of turbulent flow features on mass-transfer process is investigated using the available data. These data demonstrated a significant influence of the friction velocity u^* on solutes flux for both data sets supporting the assumption that vortices in the near-wall region would affect that flux.

Keywords: Environmental hydraulics, sediment-water interface, diffusive transport, sediment oxygen demand.

1. INTRODUCTION

The *benthic boundary layer* (BBL), sometimes termed as *bottom boundary layer*, is a zone of paramount importance to the biology, chemistry, geology and physics of the oceans, seas, lakes and even rivers. It is formed by those portions of sediment column and water column that are affected directly in the distribution of their properties and processes by the presence of the sediment-water interface. Its importance is twofold (Lorke et al., [2003]). First, within the BBL hydrodynamic energy is dissipated due to the bottom friction. Second, the BBL controls the exchange of solutes and particles between the sediment and the water. In fact, the bed could contain various types of chemicals, such as dissolved oxygen (DO), ammonia, hydrogen sulphide, organic chemical, heavy metals and radionuclides. Such chemicals can be present within

the bed both in dissolved or particulate form, i.e. attached on the particles forming the bottom sediments. Thus, chemicals sorbed onto sediments particles can be exchanged with the overlying water column through settling and resuspension or scour processes that are also greatly affected by the hydrodynamics of water flow (Chapra, [1997]). Dissolved chemicals could be then exchanged between the pore water and the water column across the sediment-water interface through mass-transfer processes, which are basically diffusive processes. Particularly, mass-transfer of dissolved oxygen, nitrogen and inorganic ions is of paramount importance in water quality and waste allocation load problems. Therefore, mass-transfer modeling can contribute to assess changes in water quality of river and streams due to the anthropic activities. The objective of the present paper is to investigate the interaction between flow hydrodynamics and fluxes of dissolved substances across the sediment-

water interface. Therefore, some literature predictive models are considered and compared. Also, the influence of turbulent flow features on mass-transfer process is investigated using experimental laboratory data collected both in flumes and benthic chambers.

2. DIFFUSIONAL EXCHANGE AT SEDIMENT-WATER INTERFACE

In a lake the transition from the background flow, far away from lake bottom, to the flow at the sediment-water interface is relatively simple due to the presence of the rigid boundary. The temporal structure of the BBL is steady as the bottom friction tends to remove fluctuations (Wüest and Lorke, [2003]). A logarithmic profile structure holds, the turbulence field is also stationary and the TKE equation is the balance between the production by Reynolds stresses and the dissipation ε , which provides the measure for turbulence level as:

$$\varepsilon = \frac{u^{*3}}{kz} \quad (1)$$

where u^* is the friction velocity [$L \cdot T^{-1}$] that is equal to $u^* = (\tau_b/\rho)^{0.5}$, k is Von Kármán constant $k=0.41$, z is water height above the sediment-water interface [L], τ_b is the bottom shear stress [$N \cdot L^{-2}$] and, ρ is water density [$M \cdot L^{-3}$].

The vertical mass-transport within the turbulent BBL is a combination of molecular and turbulent diffusion and the vertical diffusivity K_v [$L^2 \cdot T^{-1}$] is the sum of molecular D_m [$L^2 \cdot T^{-1}$] and turbulent eddy diffusivity D_t [$L^2 \cdot T^{-1}$] which depends on the dissipation rate of turbulent kinetic energy and on the stability of the density stratification (Lorke et al., [2003]). In the natural environment, typically it is $D_t \gg D_m$. However, D_t decreases steeply with the water height z . In fact as turbulent eddies approach the sediment-water interface, this interface tends to damp them as they approach closer than their length scale. Therefore, in the external area of the BBL where eddies move randomly the mass-transport is dominated by eddy diffusion, whereas moving to the sediment the influence of turbulent eddy diffusivity decreases and close to the sediment, where turbulence is low, the vertical transport is dominated by molecular diffusion. D_m values depend mainly on the solutes exchanged and on the water temperature. Also, classical boundary layer theory states that near the bottom there is a sublayer, termed viscous boundary layer (VBL), where the flow is laminar and velocity gradient is constant. This sublayer acts as a region of resistance to the transfer of momentum,

heat and mass. Within the VBL the momentum transfer is dominated by viscous forces and its thickness δ_v could be defined as the height where D_t is equal to the kinematic viscosity ν [$L^2 \cdot T^{-1}$]. The height δ_v could be estimated as:

$$\delta_v = \frac{11 \nu}{u^*} \quad (2)$$

and δ_v is typically $\delta_v \approx 10^{-2}$ m. Approaching further to the sediment-water interface, turbulent diffusivity D_t decreases up to molecular diffusivity D_m . This defines the thickness δ_c of the concentration boundary layer or diffusivity boundary layer (DBL), where the transport due to the eddies becomes negligible compared to molecular diffusion. The diffusive boundary layer is extremely thin, much smaller than the VBL. According to the dependence with z of D_t , the thickness δ_c could be related with the thickness δ_v as:

$$\delta_c = \frac{\delta_v}{Sc^\alpha} \quad (3)$$

where $Sc = \nu/D_m$ is the Schmidt number, ratio of the kinematic viscosity ν to molecular diffusivity D_m and α is a coefficient which is usually assumed to be between $1/3$ and $1/4$ (Wüest and Lorke, [2003]). Eq. (3) demonstrates that δ_c is solute-specific and is slightly temperature-dependent, as both ν and D_m change with temperature. If $\alpha=1/3$, eq. (3) shows that δ_c is for the substances of environmental concern range from $1/13$ to $1/6$ the thickness of the velocity boundary layer δ_v . Sometimes, since $Sc \approx 10^3$ and $Sc^{1/3} \approx 10$, δ_c is approximated as $\delta_c = 0.1 \cdot \delta_v$. Lorke has proposed a different scaling for δ_c , assuming that for low-energetic systems, such as lakes and reservoirs, the DBL thickness is forced by the BBL turbulence, whereas for high-energetic systems, such as streams and estuaries, δ_c value is controlled by the current velocity (Lorke et al., [2003]). Therefore, the Batchelor length scale L_B , which describes the smallest length of turbulent concentration fluctuations before molecular diffusion smoothes the remaining gradients, could be used at least as first-order approximation to define δ_c as $\delta_c = L_B$. This assumption leads to $\alpha=1/2$.

However, the thickness of DBL could not be exactly defined from a physical point of view because its boundaries are not sharp (Wüest and Lorke, [2003]). At the upper boundary a transition zone where D_t and D_m are comparable, exists, while the lower limit, i.e. the sediment-water interface, is not an horizontal plate but is sculptured into elaborate landscapes when viewed at the scale of the DBL [Røy et al.,

[2002]). Also, this interface is quite permeable so horizontal currents could diffuse slightly into the porewater. Moreover, in shallow waters heating of the sediments can lead to buoyant porewater convection, which further increases the exchange between the sediments and the water (Wüest and Lorke, [2003]). In these conditions, advection dominates on diffusive transport. Finally, in eutrophic waters, advective transport across the sediment-water interface occurs mainly by methane and carbon dioxide bubbles formed in the anoxic layer of the sediments (Wüest and Lorke, [2003]). Nevertheless if the sediment-water interface is treated as an infinite plane crossed by one-dimensional chemical gradient, the classical Fickian diffusion model could be applied. Using this approach, if D_m is constant, the mass flux across the sediment-water interface could be modeled as:

$$J_{flux} = D_m \frac{dC}{dz} \quad (4)$$

where J is the vertical mass flux per unit interfacial area [$M \cdot L^{-2} \cdot T^{-1}$] and dC/dz is the concentration gradient over z of the exchanged solute, which is the *driving force* of the diffusional process. Assuming that there is no solute production or consumption within the DBL, a linear solute concentration profile exists and Eq. (4) could be approximated as:

$$J_{flux} = D_m \frac{\Delta C}{\delta_c} \quad (5)$$

where ΔC is the $\Delta C = C_\infty - C_0$, if C_∞ and C_0 are solute concentration in the bulk water and at the sediment-water interface, respectively [$M \cdot L^{-3}$]. The ratio D_m/δ_c is usually replaced with a *conductance* term, i.e. the mass-transfer coefficient K_{m-t} [$L \cdot T^{-1}$], that relates the driving force to the mass flux. Thus, eq. (5) yields:

$$J_{flux} = K_{m-t} \Delta C \quad (6)$$

The concentrations within the DBL could be measured using microelectrodes, that allows to work at very high spatial distribution (Güss, [1998]; Lorke et al., [2003]), whereas K_{m-t} should be estimated. Generally, K_{m-t} is a function of the fluid and solute properties, surface geometry, and flow conditions (Steinberger and Hondzo, [1999]).

The diffusional transfer of solutes through the BBL influences a number of important biological and geochemical processes in the upper sediments such as the dissolution of calcium carbonate, the oxidation of organic matter and metals (iron, manganese, etc.), the removal of reactive nitrogen by denitrification, the supply of oxygen to obligate-aerobic sediment-

dwelling organisms, the growth of microbial mats, and the release of contaminants from polluted sediments (Wüest and Lorke, [2003]). Particularly, diffusional flux of dissolved oxygen (DO) towards the bed sediments from the overlying water column has been intensively investigated because it is often responsible for low and unacceptable levels of DO in the ecosystem. This diffusional flux is due to the production of oxygen-consuming substances, such as methane and ammonium ion, that are then oxidated in the aerobic layer of the bed sediments resulting in a sediment oxygen demand (SOD) (Chapra, [1997]; Gualtieri, [2001]). SOD value could be directly measured or predicted using modeling framework (Chapra, [1997]). SOD value depends both on sediment and flow over sediment characteristics (Nakamura, [1994]; Nakamura and Stefan, [1994]; Mackenthun and Stefan, [1994]; Mackenthun and Stefan, [1998]; Josiam and Stefan, [1999]; Higashino et al., [2003]; Gualtieri, [in press]). Laboratory measurements revealed a significant decrease of δ_c for increasing flow velocities (Gundersen and Jørgensen, [1990]; Hondzo, [1998]; Steinberger and Hondzo, [1999]). Thus, at low flow velocities, diffusive transport is the limiting factor of SOD production, which increases as flow velocity erodes the DBL. In this case, when near-bottom velocity is the key parameter, the process is termed as *water-side controlled*. As the velocity grows, at some point, the rate of metabolic and chemical reactions are not limited by the rate of transport of DO through the DBL. Therefore, biochemical reactions within the sediments becomes the limiting factor and SOD is independent of the current velocity over the sediment. In this case, mass-transfer is termed as *sediment-side controlled*.

In lakes, where turbulence is often low and there is an high availability of organic matter within the sediment, the microbiological activity and the consequent SOD flux is usually limited by the physical constraints of the diffusional transport, i.e. it is water-side controlled. In the present paper this case is considered and investigated.

3. PREDICTIVE MODELS FOR MASS-TRANSFER COEFFICIENT K_{m-t}

Several predictive equations have been proposed to estimate mass-transfer coefficient K_{m-t} . In this section 4 equations are presented and applied. Nakamura (Nakamura, [1994]), applying the similarity theory of the bottom shear stress and the turbulent heat o mass transfer, derived:

$$K_{m-t} = \frac{2}{\pi} \cdot n_1 \cdot \sqrt{\lambda} \cdot U \cdot Sc^{-0.75} \quad (7)$$

where $n_1=0.109$ and λ is the classic friction factor of Darcy-Weisbach equation, that is $\lambda = 8 \cdot \tau_0 / \rho \cdot U^2$, if U is the mean velocity of the flow over the sediment [L·T⁻¹]. By applying the analysis of heat transfer to the diffusional mass transfer through the diffusive boundary layer, predictive equation for K_{m-t} has been obtained, if $n_2=0.1$, as (Higashino and Kanda, 1999):

$$K_{m-t} = \frac{3 \cdot \sqrt{6}}{8 \cdot \pi} \cdot n_2 \cdot \sqrt{\lambda} \cdot U \cdot Sc^{-0.66} \quad (8)$$

Hondzo (Hondzo, [1998]) has conducted laboratory experiments to elucidate dissolved oxygen transfer mechanism at the sediment-water interface over a smooth bed. Hondzo has derived:

$$K_{m-t} = 0.0558 \cdot u^* \cdot Sc^{-2/3} \quad (9)$$

A different equation was derived from the same data set, plotting the data of dimensionless K_{m-t} against Reynolds number Re for the mean velocity U (Steinberger and Hondzo, [1999]):

$$K_{m-t} = (0.012 \pm A1) \cdot \frac{D_m}{h} \cdot Re^{0.89 \pm A2} \cdot Sc^{1/3} \quad (10)$$

where $A1 = \pm 0.001$ and $A2 = \pm 0.05$ are the 90% confidence intervals for the mean coefficient and the mean exponent, respectively.

4. COMPARISON OF AVAILABLE MODELS

In this section, the previously outlined predictive models for K_{m-t} have been tested for an idealized scenario to be compared. Assuming a channel reach with bed slope $J_b=0.001$, mass-transfer coefficient for dissolved oxygen K_{m-t} through the sediment-water interface has been computed using eqs. 7/8/9 and 10 as a function of streamflow mean velocity U . The friction factor λ has been estimated using well-known Blasius equation for smooth surfaces as $\lambda=0.316 \cdot R^{-0.25}$. All dissolved oxygen and water parameters involved in the estimation are in Table 1. Test results are presented in Fig.1a, where the mass-transfer coefficient K_{m-t} has been plotted against streamflow mean velocity U . The considered range for U is from 0 to 0.40 m/s. For the Steinberger-Hondzo equation, three couples of values for A_1 and A_2 have been considered. They are termed as $S-H_{max}$, $S-H$ and $S-H_{min}$, respectively.

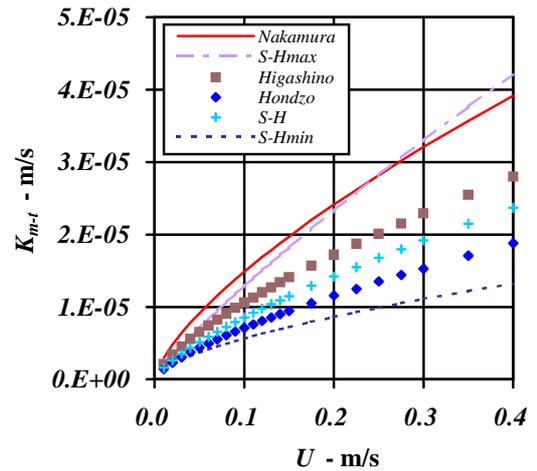
Inspection of results shows that for $U=0.40$ m/s $S-H_{max}$ and $S-H_{min}$ exhibit the highest and lowest values

for K_{m-t} . Thus, eq. (10), that was experimentally derived, encompasses all the remaining equations.

Table 1 – Input data for idealized scenario

Channel	
Slope J_b	0.001
Dissolved oxygen	
Molecular weight M – g/mole	32
Molecular diffusivity D_m – m ² /s	1.80×10^{-9}
Schmidt number Sc	557.22
Water	
Temperature T – °C	20
Density ρ – kg/m ³	998.15
Specific weight γ – N/m ³	9787.89
Surface tension T_s – N/m	0.07276
Kinematic viscosity ν – m ² /s	1.003×10^{-6}

Fig.1a - K_{m-t} vs U



The range of mass-transfer coefficient K_{m-t} for $U=0.40$ m/s is comprised from 1.32×10^{-5} m/s to 4.22×10^{-5} m/s. Also, discarding value from $S-H_{min}$, for $U=0.40$ m/s, K_{m-t} is in the range from 1.88×10^{-5} m/s to 4.22×10^{-5} m/s. Moreover, eq. (7) and $S-H_{max}$ always exhibit similar result. Finally, Higashino eq. (8) provides intermediate K_{m-t} values.

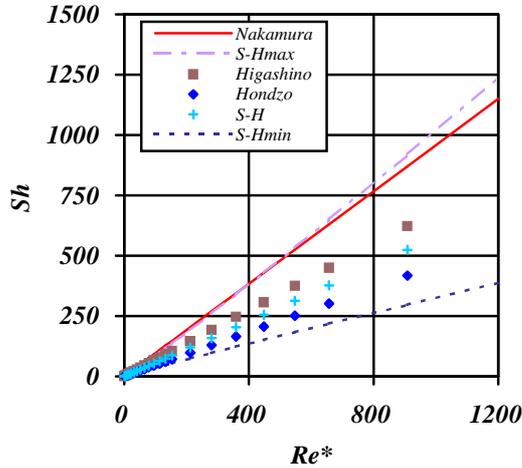
Fig.1b presents test results in terms of Sherwood number Sh against Reynolds number Re^* based on the friction velocity u^* . The Sherwood number is:

$$Sh = \frac{K_{m-t} \cdot z}{D_m} \quad (11)$$

and it represent a dimensionless mass-transfer flux. Sh values are in the range from 388 to 1242 for

Reynolds number $Re^*=1200$. The $S-H_{max}$ equation and Nakamura eq. (7) closely agree.

Fig.1b - Sh vs Re^*



5. ANALYSIS OF EXPERIMENTAL DATA. DISCUSSION

In this section, experimental data are analyzed to confirm the influence of friction velocity u^* on the mass-transfer process. Available data refer to SOD studies (Glud et al., [1995]; Mackenthun and Stefan, [1998]; Steinberger and Hondzo, [1999]; House, [2003]; Røy et al., [2004]; Tengberg et al., [2004]). Mackenthun-Stefan, Hondzo, House and Røy data sets were collected in laboratory flumes, whereas Glud and Tengberg data sets refer to benthic chamber measurements. Particularly, Tengberg data were collected in three different types of benthic chambers (Tengberg et al., [2004]). The temperature of the considered data is in the range from 8.6 to 23.8 °C and the Schmidt number Sc values are accordingly from 1032 to 454.

The values of Sh are in the range from 75 to 2500, while Re^* values are comprised from 145 to 1645. These values correspond to mass-transfer rate K_{m-t} values from 1.74×10^{-6} to 3.88×10^{-5} m/s and friction velocity u^* values in the range from 0.0018 to 0.0190 m/s. Near-bottom current speed are in lakes usually in the range from 0.02 to 0.1 m/s, but they can reach also more than 0.2 m/s during storms, especially in shallow waters (Wüest and Lorke, [2003]). In coastal areas the flow velocity near the bed could be typically of 0.02-0.04 m/s (Glud et al., [1995]). Thus, friction velocity u^* typically ranges from 0.0005 to 0.005 m/s in lakes (Wüest A. and

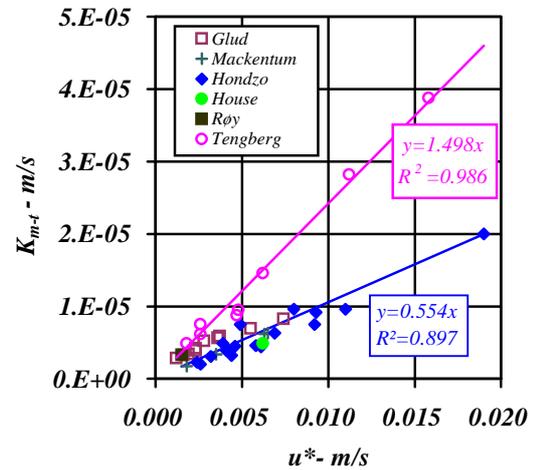
Lorke A., [2003]), whereas in streams and river u^* belongs to the range from 0.0001 to 0.01 m/s (Higashino et al., [2004]).

The experimental data are presented in Fig.2, where K_{m-t} is plotted against u^* . Notably, friction velocity u^* data were not available in the Glud data set. They were calculated using the equation proposed by Pullin et al. (Pullin et al.) for radial flow impellers:

$$u^* = 0.026 N D \frac{B}{h} \quad (12)$$

where N is the impeller rate of rotation [T^{-1}], D is the impeller diameter [L], B is the width or diameter for the chamber [L] and h is the height from the sediments to the stirrer [L].

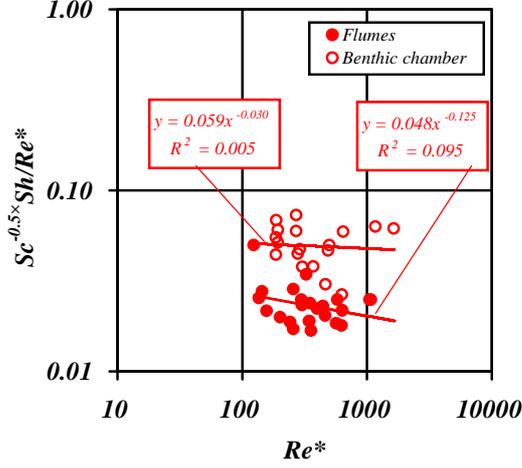
Fig.2 - Experimental data



The analysis of data generally confirms the dependence of mass-transfer rate K_{m-t} from the friction velocity u^* . A linear relationship between K_{m-t} and u^* is supported both for flumes and benthic chambers data (Fig.2). Data collected in benthic chambers are generally higher than those taken in laboratory flumes. However, u^* Tengberg data were obtained through an hydrodynamic characterization of the chambers where a PVC plate simulated the sediments surface. Thus, u^* values only give indications about the prevailing hydrodynamic conditions in the chambers during the sediment incubation (Tengberg et al., [2004]). Since sediments surface is rougher than PVC plate, u^* and also Re^* were underestimated and the slope of Tengberg data in Fig.2 should be lower. As Tengberg data, Hondzo data set also refers to hydraulically smooth bed of artificial or riverine sediments (Hondzo, [1998]).

Glud data were collected on a rough sediment surface (Glud et al., [1995]).

Fig.3 - LE and SE models



To better understand how hydrodynamics control the rate of mass-transfer turbulent flow features, such as turbulent eddies size, should be considered. In a turbulent flow the length scale of the eddies ranges from the flow domain, i.e. integral scale eddies, to smaller sizes, i.e. Kolmogorov scale eddies. At the integral scale, larger eddies break down into multiple smaller eddies efficiently transferring their energy with little loss. At the Kolmogorov scale, viscosity converts kinetic energy into heat (Pope, 2000). Thus, it is possible to assume that mass-transfer process at sediment-water interface would be controlled either by larger eddies either by smaller eddies. This hypothesis leads to the *large-eddy* and *small-eddy* models, respectively. These models could be generally represented as:

$$\frac{K_{m-t}}{u^*} = c_1 Sc^{-0.5} Re^{*n} \quad (13a)$$

or as:

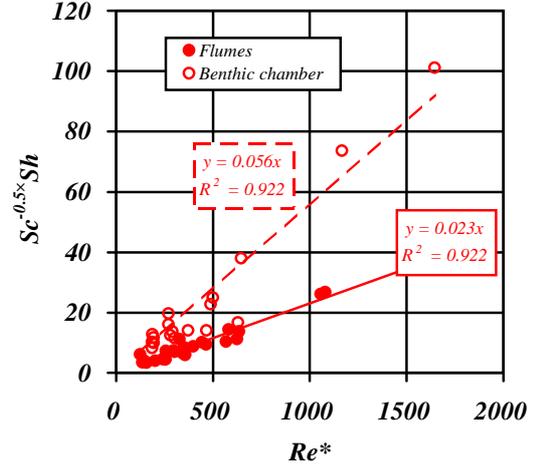
$$Sh = c_1 Sc^{0.5} Re^{*n+1} \quad (13b)$$

where c_1 is a constant, $n=-0.50$ for the *large-eddy* model and $n=-0.25$ for the *small-eddy* model.

Available experimental data are also compared with results from *large-eddy* (LE) and *small eddy* (SE) models (Fig.3). Both models are not supported since n values are -0.125 and -0.03 for flumes data and chambers data, respectively. This result is consistent with that was found by Hondzo (Hondzo, [1998]). Notably, experimental data for the mass-transfer

process at the air-water interface demonstrated that the *small-eddy* model should be preferred to *large-eddy* model (Moog and Jirka, [1999]; Gualtieri and Gualtieri [2004]).

Fig.4 - Streamwise vortex model



Counter-rotating quasi-streamwise vortices are often present at the sediment-water interface (Nino and Garcia, [1996]). Those vortices would control mass-transfer across that interface through a mechanism described by Hondzo (Hondzo, [1998]). As a turbulent motion reach the interface, it renews it to the bulk water concentration. After that, molecular diffusion returns the sediment-water interface to bed concentration. The presence of the streamwise vortex creates a pumping effect that produces ejection of low-momentum fluid, with low DO concentration on one side of the vortex core and the injection of high-momentum fluid, with high DO concentration toward the bed on the other (Hondzo, [1998]). This mechanism appears to be common to the flow over both smooth and rough surfaces (Nino and Garcia, [1996]). This vortex has a velocity scale U_v of u^* , a length scale L_v of v/u^* and a time scale T_v of v/u^{*2} . Thus, the distance of diffusive transport is given by $(D_m \times T_v)^{0.5}$ and the mass-transfer coefficient K_{m-t} is :

$$K_{m-t} \propto \frac{(D_m \times T_v)^{0.5}}{T_v} \approx \frac{D_m^{0.5}}{T_v^{0.5}} \quad (14a)$$

which yields:

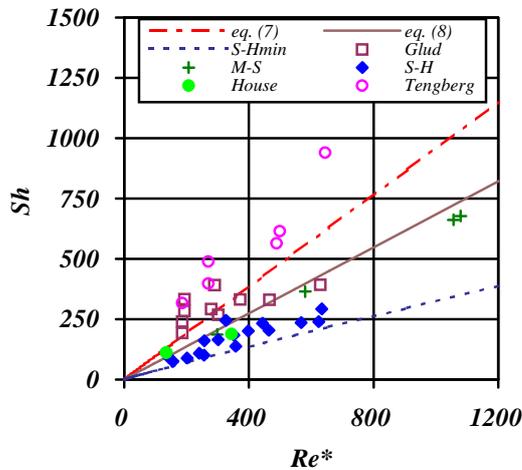
$$K_{m-t} \propto \frac{D_m^{0.5} u^*}{v^{0.5}} \approx u^* Sc^{-0.5} \quad (14b)$$

Therefore, the *streamwise vortex* (SV) model is:

$$Sh = c_2 Re^* Sc^{0.5} \quad (15)$$

where c_2 is a numerical constant. This model was also tested using the 2 sets of available experimental data (Fig.4). The analysis of data demonstrates a significant influence of the friction velocity u^* on the dimensionless flux, i.e. Sherwood number Sh , supporting the assumption that counter-rotating quasi-streamwise vortices which are present in the near-wall region could affect mass-transfer process through the sediment-water interface.

Fig.5 - Sh vs Re^*

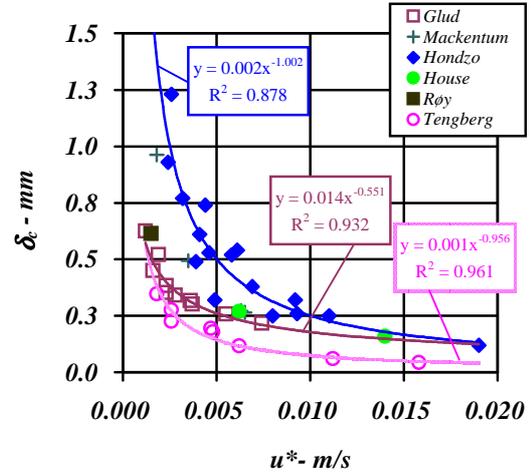


Moreover, available experimental data are compared with eqs. (7) (8) and with $S-H_{min}$ equation (Fig.5). The comparison shows that eq. (7) and $S-H_{min}$ equation encompass all the flumes data. However, most of the benthic chambers data are higher than the values predicted by the considered models. Finally, the influence of the friction velocity u^* on the thickness of diffusive boundary layer δ_c was investigated (Fig.6). The data confirmed δ_c erosion with the increasing friction velocity u^* . Particularly, regression analysis of both Hondzo and Tengberg data, which were collected on a hydraulically smooth surface, provides the equation:

$$\delta_c = c_3 u^{*p} \quad (16)$$

where c_3 is a numerical constant and p is the power law exponent, which is -1.002 and -0.956 for Hondzo set and Tengberg set, respectively. Since the exponent p is very close to 1 , these data sets confirm the linear relationship predicted by eqs. (2) and (3). However, Glud data, collected on rough surface, provide a p value of -0.551 .

Fig.6 - δ_c vs u^*



6. CONCLUDING REMARKS

Modeling the fluxes of solutes across the sediment-water interface is a relevant contribute water quality analysis of rivers and lakes. The objective of the present paper was to investigate the interaction between flow hydrodynamics and dimensionless fluxes of dissolved substances across the sediment-water interface. Therefore, some literature predictive models were compared with experimental laboratory data collected both in flumes and benthic chambers. These models encompass only flume data, whereas benthic chambers data exhibit higher values. Also, the influence of turbulent eddies on mass-transfer process was investigated using the available data. These data demonstrated that experimental data collected both in flumes and in benthic chambers are significantly correlated with flow friction velocity u^* supporting the hypothesis that counter-rotating streamwise vortices which are present in the near-bed region could affect mass-transfer process through the sediment-water interface.

REFERENCES

- Chapra S.C. (1997). *Surface water quality modeling*. McGraw-Hill, New-York
- Glud R.N., Gundersen J.K., Revsbech N.P., Jørgensen B.B. and H, Hüttl M. (1995). Calibration and performance of the stirred flux chamber from the benthic lander Elinor. *Deep Sea Research I*, vol.42, n.6, pp.1029-1042

- Gualtieri C. (2001). Dimensionless steady-state NSOD model. *Proceedings of 29th IAHR Congress*, Beijing, China, September 16/21
- Gualtieri C. and Gualtieri P. (2004). Turbulence-based models for gas transfer analysis with channel shape factor. *Environmental Fluid Mechanics*, vol.4, n.3, September 2004, pp.249-271
- Gualtieri C. (in press). Discussion on “M.Higashino, H.G.Stefan and C.J.Gantzer: Periodic diffusional mass transfer near sediment/water interface: Theory. *J.Env.Eng., ASCE*, vol.129, n.5, May 2003, pp.447-455.” in press
- Higashino M. and Kanda T. (1999). Fundamental studies on release of dissolved substance from bottom sediment to flowing water. *Proceedings of 28th IAHR Congress*, Graz, Austria, August 22/27, 1999
- Higashino M., Stefan H.G. and Gantzer C.J. (2003). Periodic diffusional mass transfer near sediment/water interface: Theory. *J.Env.Eng., ASCE*, vol.129, n.5, May 2003, pp.447-455
- Higashino M., Gantzer C.J. and Stefan H.G. (2004). Unsteady diffusional mass transfer at the sediment/water interface: Theory and significance for SOD measurements. *Water Research*, vol.38, pp.1-12
- Hondzo M. (1998). Dissolved oxygen transfer at the sediment-water interface in a turbulent flow. *Water Resources Research*, vol.34, 12, pp.3525-3533
- House W.A. (2003). Factors influencing the extent and development of the oxic zone in sediments. *Biogeochemistry*, vol.63, pp.317-333
- Josiam R.M. and Stefan H.G. (1999). Effect of flow velocity on sediment oxygen demand: comparison of theory and experiments. *J.American Water Resources Association*, vol.35, n.2, pp.433-439
- Lorke A., Müller B., Maerki M. and Wüest A. (2003). Breathing sediments: The control of diffusive transport across the sediment-water interface by periodic boundary-layer turbulence. *Limnology and Oceanography*, vol.48, n.6, pp.2077-2085
- Moog D.B. and Jirka G.H. (1999). Air-water gas transfer in uniform channel flow. *J.Hydraulic Engineering, ASCE*, 125, 1, January 1999, pp.3-10
- Mackenthun A.A. and Stefan H.G. (1994). Experimental study of sedimentary oxygen demand in lakes; dependence on near-bottom flow velocities and sediment properties. *University of Minnesota, St.Anthony Falls Hydraulic Laboratory, Project Report n.358*, Minneapolis, MI
- Mackenthun A.A. and Stefan H.G. (1998). Effect of flow velocity on SOD: experiments. *J.Env.Eng.Div. ASCE*, vol.124, n.3, pp.222-230
- Nakamura Y. and Stefan H.G. (1994). Effect of flow velocity on SOD: theory. *J.Env.Eng.Div. ASCE*, vol.120, n.5, pp.996-1016
- Nakamura Y. (1994). Effect of flow velocity on phosphate release from sediment. *Water Science & Technology*, vol.30, n.10, pp.263-272
- Nino Y. and Garcia M.H. (1996). Experiments on particle-turbulence interactions in the near-wall region of an open channel flow: Implications for sediments transport. *J. Fluid Mechanics*, vol.326, pp.285-319
- Pullin C., Oldham C. and Ivey G. . A simple estimation of friction velocities in stirred benthic chambers. Centre for Water Research Report ED 1691CP
- Røy H, Hüttel M. and Jørgensen B.B. (2002). The role of small-scale sediment topography for oxygen flux across the diffusive boundary layer. *Limnology and Oceanography*, vol.47, n.3, pp.837-847
- Røy H, Hüttel M. and Jørgensen B.B. (2004). Transmission of oxygen concentration fluctuations through the diffusive boundary layer overlying aquatic sediments. *Limnology and Oceanography*, vol.49, n.3, pp.686-692
- Steinberger N. and Hondzo M. (1999). Diffusional mass transfer at sediment-water interface. *J.Env.Eng.Div. ASCE*, vol.125, n.2, pp.192-200
- Tengberg A., Stahl H., Gust G., Müller V., Arning U., Andersson and Hall P.O.J. (2004). Intercalibration of benthic flux chambers I. Accuracy of flux measurements and influence of chamber hydrodynamics. *Progress in Oceanography*, vol.60, pp.1-28
- Wüest A. and Lorke A. (2003). Small-scale hydrodynamics in lakes. *Annual Review in Fluid Mechanics*, vol.35, pp.373-412