CHAPTER 218

Turbulence and mud sedimentation: A Reynolds stress model and a two-phase flow model

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Abstract

Two sophisticated research models, previously used and validated in internal flows, are now applied to sediment laden flows: they give a thorough insight into the vertical distribution of sediment concentration, flow-sediment interaction, stratification and inhibition of vertical mixing. The Reynolds stress model is well adapted to investigate stratification due to density effects: it gives information on turbulent fluxes of momentum and concentration, eddy diffusivity and eddy viscosity profiles, reduction of the bottom shear stress due to the presence of sediment. Results show that suspended sediments affect turbulence even at low concentrations of 1g/l. A dimensional analysis seems to indicate that these stratification effects are not well accounted for in laboratory experiments.

The two-phase flow model enables to describe the vertical profile of sediment in its continuity, from the water surface down to/through the bed, without any definition of the bed water interface. Cohesive sediment processes such as deposition, erosion or consolidation are treated as flow-particles interactions: thus, the model helps in identifying the governing parameters -floc size and density, effective stress- and does not require the classical empirical laws commonly used to describe these processes.

Introduction

State of the art for simulation of cohesive sediment transport has been considerably enhanced in the past ten years but remained tied to the knowledge of physical processes (Teisson, 1991). The sink and sources terms near the bed (Parker, 1986) and the induced repartition of sediment throughout the water column contribute to the budget of sediment, and consequently to the final output of the models. A better understanding of the complex dynamics of the vertical structure of cohesive sediment suspension (Mehta, 1989a&b) is therefore required to improve the predictability of the models.

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In that context, it is well known that turbulence is the factor responsible for maintaining in suspension very fine sediment, but is surprisingly still crudely represented in usual cohesive sediment transport models.

Turbulence in sediment laden flows has been most often studied for the case of non cohesive sediment (Lyn, 1986). The marked effects of the presence of sediment upon the flow has been highlighted by numerous laboratory experiments, as reviewed and re-analysed by Mac Lean (1991) or Villaret and Trowbridge (1991). In the experiments, discussion in general focused on the deformation of the vertical profile of velocity, whereas less attention was paid on the concentration profile by itself.

For cohesive sediment, Gust (1976) observed a turbulent drag reduction for flows over cohesive bed, even for a dilute suspension. Krone (1986) outlined the role played by aggregates, which so increase the volume concentration of suspended material that it affects the flow at even modest weight concentrations.

Field observations displayed in numeral situations that stratification effects in coastal and estuarine areas might be unrelated to thermohaline effects, but well due to the presence of sediment, since highly stratified vertical sediment concentration profiles commonly occur in otherwise vertically mixed flow regimes (Metha, 1989a).

The way that sediment affects turbulence which in turns controls deposition or erosion is of prime importance: large deposition rate at slack waters, generation of fluid mud layers, reduction of the bottom shear stress by the presence of sediment are still unsolved problems.

Two sophisticated research models, previously used and validated in internal flows, have therefore tentatively been applied to sediment laden flows (§ 1 and 2): they give a thorough insight into the vertical distribution of sediment concentration, flow-sediment interaction, stratification and inhibition of vertical mixing.

1. A one dimensional second order stress flux model

Most numerical models rely on the eddy viscosity (ν_t) and diffusivity (κ_t) concepts for the modelling of the turbulent stress and flux, and the influence of the sediment load on turbulence is then taken into account in various ways : this interaction can affect either only the eddy diffusivity distribution (van Rijn, 1990) or both eddy diffusivity and viscosity profiles through a (gradient) Richardson number dependency following Munk and Anderson (1948). This last approach has led to satisfactory results, most often in reproducing the behaviour of lutoclines (Wolanski et al., 1988 ; Mehta and Ross, 1989; Smith and Kirby, 1989; Costa and Mehta, 1990). Profiles for ν_t and κ_t are always derived from an assumed clear-water distribution for ν_t (parabolic or parabolic-constant profile).

However, some discrepancies between predictions and experiments or field measurements, some shortcomings (mainly in predicting the bottom friction velocity), let investigators suspect that a stronger interaction between hydrodynamics and sediment exists. This, together with the good results obtained for the atmospheric stratified boundary layer, have pleaded for the use of higher accuracy turbulence models for sediment laden flows. Successfully were applied k-ε model (Celik and Rodi, 1988), algebraic stress model (Hanjalic et al., 1982 ; Sheng and Villaret, 1989) and Reynolds stress model (Teisson et al., 1991 ; Brors, 1991). Hamm et al. (1992), using Sheng and Villaret model, investigated the influence of clear or loaded water on erosion laws in laboratory experiments: they found that stratification effects were most often negligible on bottom shear stress in laboratory experiments. Brors (1991) pointed out that, for the simulation of turbidity currents, only the Reynolds stress model (RSM) appeared to be realistic, k-ε and algebraic stress models giving in particular wrong concentration profiles.

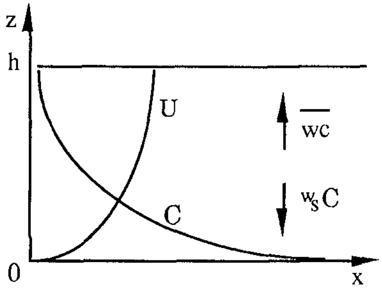
1.1 Scope of this study

The three essential processes involved here are the generation of an upward turbulent flux of particles, the damping of turbulence by gravity and the fluid-particle interaction. If we consider only small particles (mud and not sand), then neglecting the added-mass and other Basset forces, the problem reduces to the stratified turbulence problem. This is the well known thermal hydraulics or atmospheric boundary layer problem, for which Reynolds stress models have been developed and validated at LNH.

What can be expected from the RSM are, of course, mean concentration and velocity profiles, but also turbulent shear stress and sediment flux profiles, eddy diffusivity and viscosity distributions, Richardson numbers profiles and bottom friction velocity. In a more general meaning, the RSM could be helpful in understanding the mechanism of gravity effects and in evaluating the range of availability of the above mentioned turbulence models.

1.2. Mean equations

We consider a long free surface, flat bottom channel loaded with a volumetric concentration of sediments C, the settling velocity of the particles being w_S. The profile of mean velocity U(z) and mean concentration C(z) along the vertical in a channel flow are obtained from the following equations (capital letters represent mean values, small letters fluctuations and $\overline{\cdot}$ a statistical average) :



$$\frac{\partial U}{\partial t} = -\frac{1}{\rho_w}\frac{\partial P^*}{\partial x} - \frac{\partial \overline{uw}}{\partial z} + \nu \frac{\partial^2 U}{\partial z^2}$$

$$\frac{\partial C}{\partial t} + w_s \frac{\partial C}{\partial z} = -\frac{\partial \overline{wc}}{\partial z} + K \frac{\partial^2 C}{\partial z^2} \qquad (1)$$

with $w_s < 0$ (constant)

K : molecular sediment diffusivity

At equilibrium, the total shear stress ($\tau = \nu \partial U / \partial z - \overline{uw}$) is linear so that its gradient is proportional to the pressure gradient $\partial P^* / \partial x$; the settling velocity w_S induces a downward flux compensated by the gradient of turbulent concentration flux $\overline{wc}$.

1.3. Physical processes

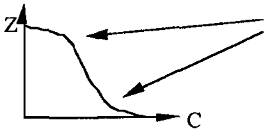
When applying a Reynolds stress model, turbulence modelling assumptions are introduced only in the transport equations of the turbulent fluxes and stresses ; so the production and stratification effects, induced on these second moments by velocity and concentration gradients, are accounted for exactly. Although it is apparently complex with the 8 equations added to the mean equations (1), the Reynolds stress model allows a good understanding of the physical processes governing the turbulence-sediment interaction. Let us here concentrate on the three equations driving turbulent phenomena, that are those for the upward turbulent flux of sediments and for the variances of the vertical and horizontal fluctuating velocities. Developing only the (exact) production terms, while the pressure-strain Φ, diffusion D and dissipation ε terms respectively are to be modelled, they read :

$$\frac{\partial \overline{wc}}{\partial t} = -\overbrace{\overline{ww}\frac{\partial C}{\partial z}}^{A} - g\beta\overline{c^2} + \Phi_{i,c} + D_{i,c} \tag{2}$$

$$\frac{\partial \overline{ww}}{\partial t} = \overbrace{-2g\beta\overline{wc}}^{B} + \Phi_{i,j} + D_{i,j} - \frac{2}{3}\varepsilon \qquad \text{with } \beta = -\frac{(\rho_s - \rho_w)}{\rho_w} \tag{3}$$

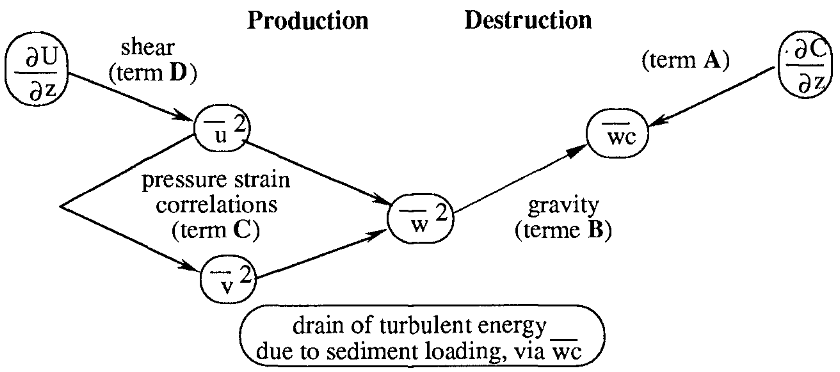
$$\frac{\partial \overline{uu}}{\partial t} = \overbrace{-2\overline{uw}\frac{\partial U}{\partial z}}^{D} + \overbrace{\Phi_{i,j}}^{C} + D_{i,j} - \frac{2}{3}\varepsilon \tag{4}$$

Considering the concentration profile allows first to understand the mechanism of turbulence destruction due to the sediment load :



The concentration gradient is large, negative here. Then term A has large positive values, and (2) leads to an increase of the turbulent flux of concentration $\overline{wc}$. In turn, term B is negative, and (3) implies that the variance ww decreases.

As a consequence, turbulence tends to become two-dimensional and energy is then drained from the two horizontal components u and v (eq. (4)), via the pressure-strain correlations, to feed w (eq. (3)). Finally, this mechanism leads to a decrease of the whole turbulence, which can be summarised by the following scheme :



This sketch naturally yields to the definition of the flux Richardson number Ri_f (the ratio of terms B/D), which is the fraction of the turbulence production that can be diverted from dissipation to act against gravity.

1.4. Reynolds stress model

The second moment closure modelling adopted here is a classical one. A wall echo term is added to take into account the redistribution of energy from the vertical component of the fluctuating velocity w at the bottom, as at the free surface following the proposition by Gibson and Rodi (1989). Anisotropy affects the dissipation equation through the Launder and Tselepidakis (1991) proposition. The model has first been assessed against clear-water open channel flow data (Nakagawa et al., 1975; Komori et al., 1982).

Results are presented here for the schematic case of a steady, plane open channel flow of depth h = 1 m, bulk velocity U_0 = 0.5 m/s (Re = 500 000), bulk sediment concentration of 1 g/l (volumetric concentration $C_0 = 3.75\ 10^{-4}$) and particle settling velocity w_s = -0.001 m/s. Initial conditions are a homogeneous sediment concentration over the water depth and a logarithmic velocity profile, boundary condition for the concentration is a zero flux condition at both bottom and free surface.

Figures 1 to 3 show the effect of coupling the gravity (setting g = 0 to g = 9.81 m/s^2) in the equations for the turbulent stresses on the mean velocity and concentration, and on the shear stress.

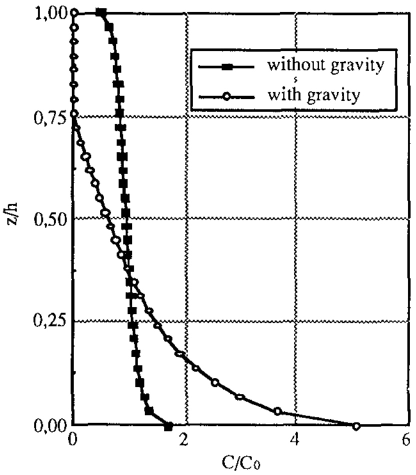


Fig. 1

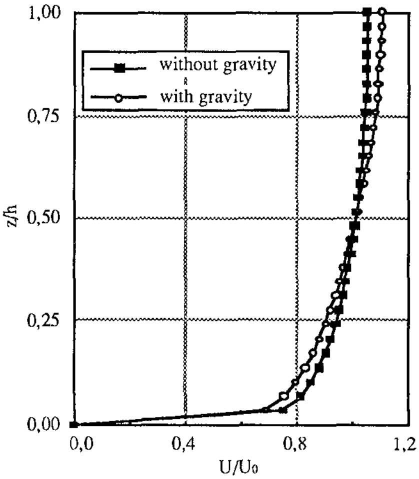


Fig. 2

This effect is seen to be quite strong for the sediment concentration considered here. The concentration dies out to 0 in the 25% upper part of the water column and is multiplied by nearly a factor 3 at the bottom (fig. 1), the velocity profile significantly deviates from the logarithmic one (fig. 2) and the bottom shear stress is reduced by more than 20% at the bottom (fig. 3).

Some experimental and field investigations have already highlighted such deviations from clear-water hydrodynamics. Gust (1976) found, from experiments with cohesive sediment in sea-water, that the bottom friction velocity u_* was reduced by 20% to 40% and that the thickness of the viscous sub layer was increased by a factor varying between 2 and 5 ; furthermore, the logarithmic profile for the velocity was shown to be no more valid in the near-wall region. Soulsby and Wainwright (1987) have also pointed out, from field measurements for non-cohesive sediments, that u_* could be over-estimated by more than 50% when using the logarithmic velocity profile, neglecting thus the suspended sediment effects.

The gradient Richardson number (fig. 4), which is usually used to introduce the influence of sediment on turbulence, is seen to be only weakly correlated to the flux Richardson number, which is the sound parameter to measure buoyancy effect according to the Reynolds stress equations. This suggests that the Reynolds analogy as well as the Munk-Anderson approach could present some shortage for this application.

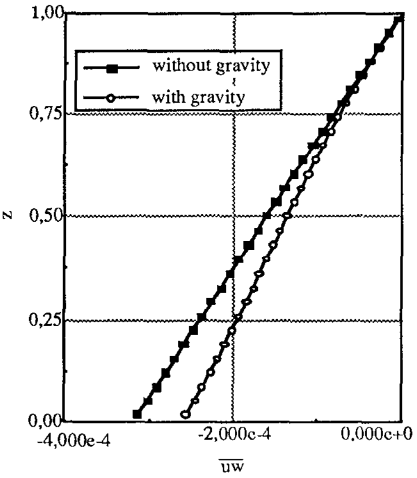


Fig. 3

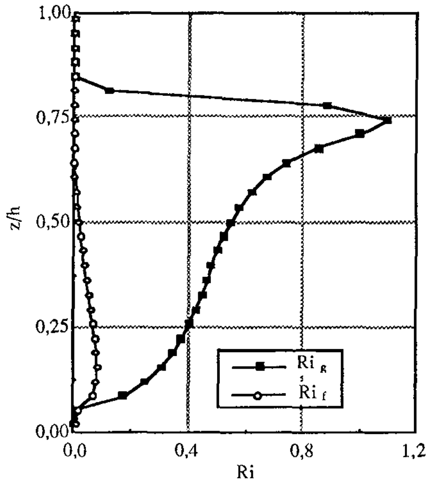


Fig. 4

1.5. Analytical developments

Direct comparison between numerical predictions and measurements is difficult for there are only few data available on cohesive sediment transport, even for the mean variables. Furthermore, as results of the RSM depend entirely on assumptions for modelling the turbulent second moments, validation should concern first turbulence predictions. But there is even less turbulence measurements... In this section we try to make some analytical developments that could help understanding RSM results.

As seen previously, the buoyancy level is defined by the flux Richardson number :

$$Ri_f = \frac{g\beta\overline{wc}}{\overline{uw}\frac{\partial U}{\partial z}} \qquad (5)$$

Assuming a logarithmic velocity profile : $\overline{uw} = -u_*^2(1-\frac{z}{h})$ and $\frac{\partial U}{\partial z} = \frac{u_*}{\kappa z}$

and when equilibrium is reached : $\overline{wc} = -w_s C$

So that, at equilibrium and under the assumption of a logarithmic velocity profile, Ri_f becomes :

$$Ri_{f\ eq} = g\beta\kappa \frac{w_s z C_{eq}}{u_*^3 (1 - z/h)} \quad (6)$$

The maximum value for Ri_f, as measured from thermal flow investigations, is 0.25 ; turbulence being totally damped by buoyancy for higher values. We can then derive a formula giving, under the above assumptions, the maximum concentration at each location in the water column that a given flow can held at equilibrium, $C_{eq\ max}$:

$$C_{eq\ max} = \frac{0.25}{g\beta\kappa} \frac{u_*^3}{w_s} \left(\frac{1}{z} - \frac{1}{h}\right) \quad (7)$$

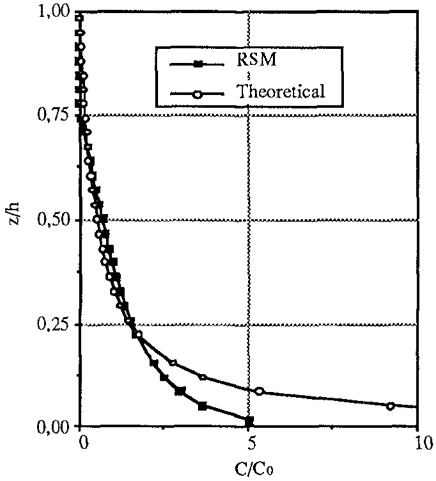


Fig. 5

This analytical expression does not give an equilibrium profile and has no physical meaning, but is an upper boundary for the sediment distribution: concentration profiles should stand below the curve drawn by the formula. According to (7), a reduction of the flow velocity by a factor 2 could reduce the maximum concentration to be carried by a factor 8.

The RSM prediction agrees well with (7) in the bulk of the flow (fig. 5), which means that the drastic reduction in sediment concentration in the upper part of the channel is due to the damping of turbulence by the sediment load. Figure 5 also indicates that the flow could hold much higher concentrations at the bottom, although (5) certainly over-estimates them there because dissipation, which is important at these locations, is neglected.

Let us now make a scale analysis on the flux Richardson number, which can be expressed, under some assumptions as seen previously, by (6). Let suppose a scale model expcriment and express by $\hat{}$ the similitude laws ($\hat{}$ = .lab / .field). Then we get the following expression for the similitude law for Ri_f :

$$\hat{Ri}_f = \frac{\hat{\beta}\ \hat{w}_s\ \hat{C}\ \hat{h}}{\hat{u}_*^3}$$

Consider now a laboratory experiment (deposition or erosion test in a flume or a carousel). Most often, the procedure is:

- to use the same mud as in the field ($\hat{\beta} = 1$ and $\hat{w}_s = 1$),
- for the same range of concentrations ($\hat{C} = 1$),

- under the same hydrodynamics conditions ($\hat{u}_* = 1$)
to derive laws to be introduced in mathematical models for prototype applications.

This procedure leads to a scaling effect on Ri_f which reduces to the vertical scale $\hat{h}$. This result, obtained here from (6) under some assumptions, has also been rigorously derived from a dimensional analysis of the Reynolds stress equations (Laurence et al., 1993). This means that, in a flume with a depth of 0.20 m, stratifications effects will be 50 times less than in the field, with a water depth of 10m, for the same range of concentrations and velocities. This could explain the weak stratification effects found by Hamm et al. (1992) in laboratory, whereas a larger influence is suspected in the field (Gust, 1978). Erosion deposition laws used in modelling high turbid environment in China (Costa and Mehta, 1990) required the use of physical parameters which were significantly different from those determined in laboratory experiments performed with local mud. They concluded that erosion-depositon fluxes can be drastically modified by sediment stratification, and that typical formulations for the bottom fluxes are believed to have limited utility in such environments. The reason could be, as stated above, that, for the same hydrodynamic forcing, in heavily laden flows, stratification effects will be far more important in the field than in the laboratory.

2. A separated two-phase flow model

Following Wallis (1969), transport of sediment in suspension can be regarded as a two-phase flow, i.e. mixture of a continuous phase (water) and a dispersed phase (mud flocs) for the case of cohesive sediment transport. And, by writing mass and momentum balance equations separately on each phase, with appropriate momentum transfer rate between phases, the two phase formulation enables to describe how the presence of particles modifies fluid flow characteristics, and reciprocally how the instantaneous fluid flow acts on particles movements.

Mass balance :

$$\frac{\partial}{\partial t} \alpha_k \rho_k + \frac{\partial}{\partial x_i} \alpha_k \rho_k U_{k,i} = \Gamma_k \tag{8}$$

where, $U_{k,i}$ is the mean velocity i-component for the continuous (k=1) and dispersed phases (k=2) respectively, α_k is the volumetric fraction, ρ_k the mean density and Γ_k the interfacial mass transfer rate between phases. In cohesive sediment transport, Γ_k accounts for water capture by mud flocs during their growth or break-up (transfer of water from the water phase to the mud flocs phase), and therefore for density change of the flocs. For the applications below, Γ_k is set to zero, and the density of the flocs remains constant.

Momentum balance :

$$\alpha_k \rho_k \frac{\partial}{\partial t} U_{k,i} + \alpha_k \rho_k U_{k,j} \frac{\partial}{\partial x_j} U_{k,i} = -\alpha_k \frac{\partial}{\partial x_i} P_1 + \alpha_k \rho_k g_i + I_{k,i} - \Gamma_k U_{k,i} - \frac{\partial}{\partial x_j}\left[\alpha_k < \rho\, u''_i\, u''_j >_k + T_{k,ij} \right] \tag{9}$$

where u''_i is the fluctuating part of the local instantaneous velocity and $< . >_k$ the averaging operator associated to phase k,

$$\alpha_k \rho_k U_{k,i} = \alpha_k < \rho\, u_i >_k \qquad < \rho\, u''_i >_k = 0$$

$< u''_{k,i}\, u''_{k,j} >_k$ is the turbulent Reynolds stress tensor of the liquid phase (k=1) and the kinetic part of the particulate stress tensor (k=2),

$T_{k,ij}$ is the molecular viscous stress tensor (k=1) and the collisional part of the particulate stress tensor (k=2), set to zero for this application.

P_1 is the mean pressure of the continuous phase,

$I_{k,i}$ is the part of the interfacial momentum transfer rate between phases which remains after substraction of the mean pressure contribution and complies with the mean jump condition derived from the local balance of momentum at the interfaces,

$$\sum_{k=1}^{2} I_{k,i} = 0$$

The closure of the averaging equations set is achieved by using :

- practical expressions in terms of the computed variables to approximate the mean interfacial transfert terms and derived from the local description of single particle transfer with the surrounding fluid ;
- second-moment modelling for the continuous phase turbulence and the kinetic (or transport) part of the particulate stress tensor ;
- constitutive relations for the mean transport properties accounting for the molecular viscous stress in the liquid phase and the collisional part of the particulate stress tensor.

Interfacial momentum transfer :

Constitutive relations for the interfacial transfer terms derive by averaging from the particulate expressions, and must be related to the mean computed variables. Neglecting the Basset force, the interfacial momentum transfer term $I_{k,i}$ induced by the relative motion of dispersed particles, can be written:

$$I_{1,i} = -I_{2,i} = \alpha_2 \rho_1 F_D V_{r,i} - \rho_1 < u''_{1,i} u''_{2,j} >_2 \frac{\partial}{\partial x_j} \alpha_2 + \alpha_2 \rho_1 C_A \left[\frac{\partial V_{r,i}}{\partial t} + U_{2,j} \frac{\partial V_{r,i}}{\partial x_j} \right] + \frac{\partial}{\partial x_j} \alpha_2 \rho_1 C_A < u''_{2,j} v''_{r,i} >_2 \qquad (10)$$

The first term on the right-hand side represents the drag force, the second the correlation between instantaneous distribution of particles and fluid pressure fluctuations, and the other ones the apparent mass force. The main part of the force induced by the fluid flow is already taken into account by the expression of the pressure gradient term in the momentum equations.

F_D, the average drag coefficient is written in terms of the local mean particle Reynolds number which accounts roughly for the flocs overcrowding and the non-linear dependence on the relative velocity fluctuations :

$$F_D = \frac{3}{4} \frac{C_D(<Re>)}{\bar{d}} < \left| \vec{v}_r \right| > \qquad < \left| \vec{v}_r \right| > \approx \sqrt{V_{r,i} V_{r,i} + < v''_{r,i} v''_{r,i} >}$$

$$C_D(<Re>) = \frac{24}{<Re>} \left[1 + 0.15 <Re>^{0.687} \right] \alpha_1^{-1.7} \qquad <Re> = \frac{\alpha_1 < \left| \vec{v}_r \right| > \bar{d}}{\nu_1} \qquad (11)$$

$V_{r,i}$, the averaged value of the local relative velocity between each particle and the surrounding fluid, can be expressed in function of the total relative mean velocity $\Delta U_i = U_{2,i} - U_{1,i}$ and a drifting velocity $V_{d,i}$ due to the correlation between the instantaneous distribution of particles and the turbulent fluid motion at large scales with respect to the particle diameter :

$$V_{r,i}=\left[U_{2,i}-U_{1,i}\right]-V_{d,i} \qquad V_{d,i}=<\tilde{u}_{1,i}>_2-U_{1,i}=<u''_{1,i}>_2 \tag{12}$$

The drifting velocity $V_{d,i}$ takes into account the dispersion effect due to the particles transport by the fluid turbulence and reduces to the single turbulent correlation between the concentration fluctuations and the turbulent fluid velocity when the particle diameter is tending towards zero with respect to the smallest turbulent length scales. According to the theoretical case of particles suspended in homogeneous turbulence (Deutsch and Simonin, 1991) the velocity $V_{d,i}$ is written as follows :

$$V_{d,i}=-D^t_{12}\left[\frac{1}{\alpha_2}\frac{\partial\alpha_2}{\partial x_i}-\frac{1}{\alpha_1}\frac{\partial\alpha_1}{\partial x_i}\right] \tag{13}$$

where the fluid-particle turbulent dispersion coefficient D^t_{12} is given in terms of the covariance between the turbulent velocity fluctuations of both phases and a fluid-particle interaction turbulent characteristic time.

The turbulence model

Turbulence is modelled throught the concept of turbulent viscosity $\nu_k{}^t$ prescribed by a standard k - ε model, with a different treatment of the continuous and dispersed phase (Bel F'Dhila and Simonin, 1992).

A further balance equation for the particle number :

$$\alpha_2\rho_2\frac{\partial}{\partial t}X_p+\alpha_2\rho_2U_{2,j}\frac{\partial}{\partial x_j}X_p=\frac{\partial}{\partial x_j}\alpha_2\rho_2D^t_2\frac{\partial}{\partial x_j}X_p-X_p\Gamma_2+\Gamma_p \tag{14}$$

where Γ_p, the rate of change in the particle number due to breakup and agglomeration.

N_p, the mean particle number by unit volume of the two-phase mixture can be expressed directly in function of the variable X_p (the mean particles number by unit of mass of the dispersed phase) :

$$N_p=\alpha_2\rho_2X_p$$

and leads to the general definition of the mean diameter :

$$\frac{\pi\bar{d}^3}{6}=\frac{1}{\rho_2X_p}$$

For cohesive sediment transport, this equation would account for diameter change of the flocs due to break-up or aggregation. For the applications below, this equation is not used , and the diameter of the flocs remains constant.

Applications

The definition of the transition from water body to bed is rather vague (Parker, 1986; Mehta, 1989a) and is one of the motivations to design an approach where this definition is circumvented. In that context, the two-phase flow model appears as the most complete model of the whole process from the water surface to the rigid bed.

Once the diameter and the density of the "inclusions" have been defined as data, the fall velocity V_r of particles is not prescribed but is an output of the model, as an exemple of flow-sediment interaction (fig. 6).

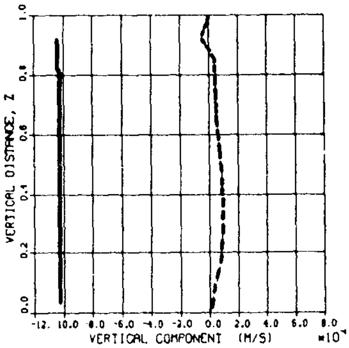


Fig.6. Mean relative velocity V_r (solid line) and mean velocity of the dispersed phase U_2 (dash line), at low concentrations.
No hindered settling (V_r constant), equilibrium conditions ($U_2 \approx 0$)

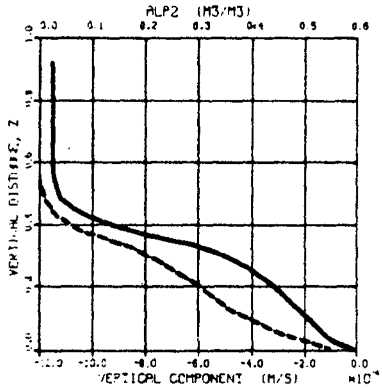


Fig. 7. Vertical profile of α_2, volumetric fraction of the solid phase (dash line), and vertical relative velocity V_r (solid line), with strong hindering effect near the bottom for large concentration

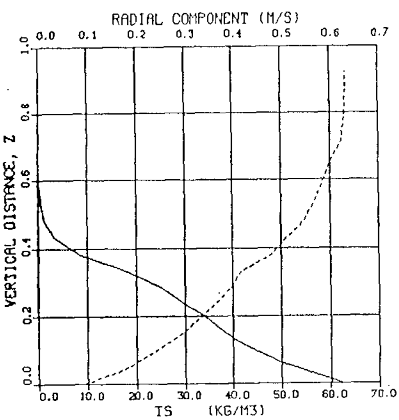


Fig.8. Vertical profile of concentration (solid line) and velocity (dash line)

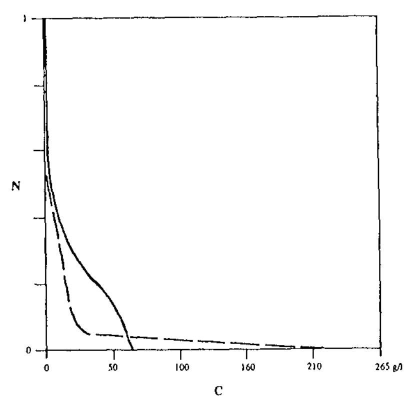


Fig. 9. Vertical profile of concentration for the same settling velocity of 1 mm/s: for flocs of density 1070 kg/m3 (solid line) and for non cohesive sediment of density 2650 kg/m3 (dashed line).

Another output of the two-phase flow model is the absolute vertical velocity U_2 of particles in the presence of turbulence: it can be oriented downwards (turbulence too weak to stand deposition), or be zero (equilibrium condition on fig.6) or even be orientated upwards (for instance when mixing in the water column during erosion in accelerating currents). This a fundamental result for sedimentation problem.

The capability of two phase flow approach to analyse hindered settling (fig. 7) has been pointed out by Thacker and Lavelle (1977).The counter flow of the fluid upwards through the falling sediment is included in the model, as a consequence of the mass balance equation (8). The presence of others particles is taken into account in the drag coefficient expresion (11).

Results with the two-phase flow model have been obtained for the same schematic case as for the Reynolds stress model (water depth 1 m; mean velocity 0.5 m/s), but for higher concentrations. Vertical profile of flow velocity and concentration have been computed in presence of mud flocs of density 1070 kg/m^3 and diameter 160 µ (settling velocity 1 mm/s after the Stokes law), with a bulk initial concentration of 0 (clear water), 4, 8 g/l.

At very low concentration (fig. 6), the profile is in equilibrium ($U_2 \approx 0$) and the relative velocity V_r given by the model corresponds to the Stokes value (-1 mm/s). For the latter case (8 g/l), classical equilibrium profiles cannot be sustained, and the initial profile completely collapses, with volumetric concentration up to 0.60 near the bottom for the final profile (expressed in massic concentration on fig. 8); hindered settling is very pronounced, at such concentrations (fig. 7). Concerning the liquid phase, the classical logarithmic profile of velocity in clear water is modified by the presence of sediment, with a kink located at the level where the concentration strongly increases downwards (fig. 8). From all the tests performed, the effect of the presence of particles on the turbulence of the flow appears to be noticeable even at "low" concentrations, the order of 1 g/l.

The role of floc size and floc density, which can be taken into account only in such models, is illustrated on fig. 9: this figure presents the vertical profile of concentration, obtained with a flow of 0.5 m/s, for an initial bulk concentration of 8 g/l, for two sediments exhibiting the same settling velocity of 1 mm/s:

- non cohesive sediment (sand) of density 2650 kg/m3 and diameter 31 µ,
- the same mud flocs as before of density 1070 kg/m3 and diameter 160 µ.

The Stokes law gives, for the two sediments a settling velocity of 1 mm/s. However, they behave completely differently: sand settles rapidly to the bed, whereas mudflocs form a mobile suspension (fig. 9). The volumetric concentration for flocs is of the order of 0.10 to 0.50, against 0.01 to 0.05 for sand, in order to have the same massic concentration: this induces hindered settling for flocs, the water has difficulty to escape between the falling flocs, as nearly half of the volume is occupied by the flocs. Thus, the two-phase flow approach points out that the floc size might play an important role in the generation of fluid mud layers; massic concentration and settling velocity are not sufficient to explain the phenomenon (they are the same for the two sediments). Floc size and density, volumetric concentration seem more appropriate than settling velocity and massic concentration.

In the near future, the two-phase flow model should offer the possibility to study aggregations and break-up of flocs as a function of the level of turbulence, through a further equation (14) on the number of flocs already included in the model, provided information is available on Γ_p.

Last but not least, Wallis (1969) first emphasized the analogy between two-phase flow and classical consolidation equations, by adding in the momentum balance equation the interparticle stress , the so-called "effective stress" in consolidation theories, which appears as $T_{k,ij}$ in (9).

Adressing the proper parameters (floc size and density, rate of aggregation and break-up, effective stress) to describe cohesive sediment processes, the two-phase flow approach should help to reduce empiricism in commonly used formulae and encourage further experimental work on these parameters

Conclusion

Today, a new generation of models, such as Reynolds stress models and two-phase flow models has become available in the industrial domain and we have attempted to apply the most recent ideas and concepts to cohesive sediment laden flow. The philosophy was to represent the hydrodynamics, turbulence and flow sediment interaction in the most accurate way, and, if possible, to get a new understanding of the physical processes in return.

Results provided by the Reynolds stress model show a strong damping of turbulence leading to significant deformation of the mean velocity, concentration and shear stress profiles, when compared with the clear-water ones. According to (6), for a given flow, stratification directly depends on the water depth, the sediment concentration and the settling velocity. The strong effects obtained here for "macroflocs" with w_s = 1 mm/s would then be reduced accordingly for settling velocities 10 or 100 times lower.

In the two-phase flow approach, the physical processes of cohesive sediment are analysed and modelled as flow sediment interaction. This model enables modelling of the vertical profile of concentration from the free surface down into the bed without any definition of the bed water interface, as cohesive sediment processes are treated internally.

Henceforth, the formalism of these turbulence models offers a rigorous framework to study in a unified way the processes of mud transport or sedimentation. Empiricism in the formula commonly used in cohesive sediment transport modelling could be reduced thanks to these better representation of flow-sediment interactions, and predictability of the models consequently increased.

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