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Initial growth of phytoplankton in turbid estuaries: A simple model

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Abstract

An idealised model is presented and analysed to gain more fundamental understanding about the dynamics of phytoplankton blooms in well-mixed, suspended sediment dominated estuaries. The model describes the behaviour of subtidal currents, suspended sediments, nutrients and phytoplankton in a channel geometry. The initial growth of phytoplankton and its spatial distribution is calculated by solving an eigenvalue problem. The growth rates depend on the position in the estuary due to along-estuary variations in nutrient concentration and suspended sediment concentration. The model yields an insight into how the onset of blooms in the model depends on physical and biological processes (turbulent mixing, fresh water discharge, light attenuation, imposed nutrient concentrations at the river and sea side). In particular, the model demonstrates that the joint action of spatial variations in turbidity and in nutrients causes the maximum phytoplankton concentrations to occur seaward of the estuarine turbidity maximum.

Keywords: Estuarine dynamics; Phytoplankton; Turbidity; Stability analysis

1. Introduction

In many estuaries characteristic (spatial and temporal) patterns of phytoplankton (or chlorophyll) concentrations are observed. Field data collected over two decades in San Fransisco Bay (Cloern, 1991) revealed that during each spring phytoplankton blooms occur and that the bloom is more intense during neap tide than during spring tide. Observations of algae in the York river (Sin et al., 1999), a tributary to the Chesapeake Bay (VA), show that during the winter–spring a strong algal bloom is often present in the mid-reach of the mesohaline zone. During the summer a smaller bloom often occurs in the transition zone from fresh water to mesohaline water.

Concepts to explain the behaviour of phytoplankton all use that phytoplankton growth is limited by light and nutrients and that decay of phytoplankton is due to respiration, zooplankton grazing and benthic grazing. It was argued by Sverdrup (1953) that blooms in the ocean occur in early spring when the surface mixed layer becomes so shallow (due to increasing heat input and reduced wind input) that algae can reach areas where sufficient light is available for them to grow. Also, vertical mixing needs to be sufficiently intense that algae can come close to the bottom where nutrient concentrations are largest.

As shown by Lucas et al. (1998), the concepts of Sverdrup cannot be straightforwardly applied to explain phytoplankton growth in coastal plain estuaries because of the different processes that are at work there. First, density stratification in estuaries is usually caused by differences in salinity, not temperature. Second, tides cause strong stirring of phytoplankton. Third, besides the bottom, the discharging river is a main source of nutrients. Fourth, local changes of nutrients and phytoplankton are also affected by horizontal transport processes (Lucas et al., 1999). Finally, light attenuation will be largely influenced by the concentration of suspended sediments in the water (May et al., 2003). The spatial and temporal distribution of suspended sediments is controlled by external forcing conditions, in particular tides and fresh water discharge

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(Burchard and Baumert, 1998) and by sediment properties (Winterwerp, 2002).

The concepts mentioned above have been incorporated into numerical models with an increasing degree of complexity (see e.g., May et al., 2003). These models have contributed considerable insight into the role of different physical and biological processes on the occurrence of phytoplankton blooms. In particular, the role of time-varying vertical mixing on the tidal and spring–neap time scale has been intensively explored.

For gaining further fundamental understanding of the results of numerical models, it is often helpful to develop and analyse idealised, semi-analytical models. Although the latter often make severe assumptions on the parameterization of processes that are accounted for, they are fast and their results can be analysed and interpreted in relatively straightforward manners. In this paper such a simple model is considered for a well-mixed, suspended sediment dominated estuary. It uses concepts that are similar to those discussed in May et al. (2003), but the focus here is on variations in currents, suspended matter and phytoplankton concentrations in a longitudinal section (from sea to river) rather than in a lateral cross-section. The model is introduced in Section 2 and it is analysed in Section 3. Results are presented and discussed in Sections 4 and 5, respectively, and finally the conclusions are given.

2. Model formulation

2.1. Domain, water motion and SSC distribution

The geometry that will be considered is that of an idealised estuary with a constant width b and constant depth h. A Cartesian coordinate system is chosen, where x, y, z are longitudinal (increasing from sea to river), lateral and vertical coordinates, respectively. Here, $z = 0$ is the undisturbed water level.

The equations describing the subtidal currents and suspended sediment concentration (SSC) are equivalent to those used in an accompanying paper (Talke et al., 2007). The flow is described by the steady, linear width-averaged shallow water equations and it is forced by an imposed fresh water discharge at the river mouth and by a horizontal density gradient due to a given salinity distribution in the channel. At the surface the stress vanishes (no wind), whilst at the bottom a no-slip condition is imposed. Salinity is assumed to be well mixed in the vertical. Using results of the above-cited study the along-channel distribution of salinity is modelled as

$$s(x) = \frac{1}{2}s_*\left[1 - \tanh\left(\frac{x - x_c}{L}\right)\right]. \tag{1}$$

Here, s_* is the salinity at sea, x_c the position at which the salinity is 50% of its value at sea and $x_c + L$ is a measure of the salt intrusion length (at $x = x_c + L$ the salinity is $0.12s_*$). The density of water, ρ, is calculated from

$$\rho(x) = \rho_0 + \beta s, \tag{2}$$

where $\rho_0(\sim 1020\,\mathrm{kg m^{-3}})$ is a constant reference density and $\beta(\sim 0.83\,\mathrm{kg m^{-3}\,psu^{-1}})$ is a coefficient. Turbidity currents induced by gradients in concentration of suspended sediments are neglected in the present model. The Boussinesq approximation is applied, i.e., variations in density are small compared to the reference density. Finally, the rigid lid assumption is made, i.e., elevations of the free surface are ignored, except in maintaining a barotropic pressure gradient.

The longitudinal velocity component $u(x,z)$ that obeys the equations of motion and boundary conditions reads (Officer, 1976)

$$u(x,z) = \frac{gh^3\beta}{48\rho_0 A_v}\frac{\mathrm{d}s}{\mathrm{d}x}\left(1 - 9\left(\frac{z}{h}\right)^2 - 8\left(\frac{z}{h}\right)^3\right) + \frac{3Q}{2bh}\left(1 - \left(\frac{z}{h}\right)^2\right), \tag{3}$$

where g is the acceleration due to gravity, $A_v(\sim 10^{-3}\,\mathrm{m^2\,s^{-1}})$ is a constant vertical eddy viscosity coefficient and $|Q|(\sim 10^2\,\mathrm{m^3\,s^{-1}})$ is the fresh water discharge. In this model Q has negative values, because the x-axis points to the upstream direction. The terms on the right-hand side describe the currents driven by the horizontal salinity gradient and by fresh water discharge, respectively. The vertical velocity component $w(x,z)$ follows from solving the continuity equation and the result is

$$w(x,z) = -\int_0^z \frac{\partial u}{\partial x}\,\mathrm{d}z'. \tag{4}$$

Together, u, w describe the classical gravitational (or estuarine) circulation (Hansen and Rattray, 1965).

Mass conservation also implies that the net volume of water transported through any cross-section is constant, i.e.,

$$\int_{-h}^{0} u(x,z')\,\mathrm{d}z' = q, \quad q = \frac{Q}{b}. \tag{5}$$

The distribution of suspended sediments is computed from the tidally averaged concentration equation. The particles are assumed to be noncohesive and have a constant settling velocity w_s $(\sim 10^{-3}\,\mathrm{ms^{-1}})$. The horizontal and vertical eddy diffusion coefficients K_h and K_v are assumed to be constant. Typical values are $K_h \sim 10^2\,\mathrm{m^2\,s^{-1}}$ and $K_v \sim 10^{-3}\,\mathrm{m^2\,s^{-1}}$. As shown in Talke et al. (2007) horizontal transport processes play an important role in maintaining an equilibrium distribution of sediment. The solution for the concentration $c(x,z)$ reads

$$c(x,z) = c_b(x) f_c(z), \quad f_c(z) = \exp(-w_s(z+h)/K_v) \tag{6}$$

and the near-bed concentration c_b follows from imposing the morphodynamic equilibrium condition (no net longitudinal transport of sediment), which was first used by Friedrichs et al. (1998). Applying this condition yields the