Modeling ice-proximal fine sediment transport associated with a subglacial buoyant jet in glacial fjord

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\section*{Abstract}
Fine sediment transport produced by a subglacial freshwater discharge is simulated with a 2D nonhydrostatic model. The circulation pattern revealed a buoyant jet issuing from the tunnel, a vertically buoyant plume and a horizontal surface plume forming part of an estuarine circulation. Jet-dominated experiments are more sensitive to the presence suspended sediment in the discharge. At low concentrations, the sediment stays in the vertical and horizontal plumes and its concentration is progressively decreased by mixing but not noticeable settling is produced through the water column. At high concentrations, the sediment settles in the far field and is transported back to the near field by the landward estuarine current. Sediment came off the surface layer through convective sedimentation, a process that was more effective than flocculation to transport sediment vertically, and showed vertical velocities faster than $1.0 \times 10^{-2}$ m s$^{-1}$. Implications of these results are discussed.

\textit{Keywords:} Glacier, convective sedimentation, plume, flocculation, buoyancy,
nonhydrostatic model, sediment dynamics, fjord
1. Introduction

Approximately one-tenth of the world coastlines are active glacimarine environments or environments where sediment is deposited after being discharged from glacier ice (Curran et al., 2004). Some of these glacimarine environments are glacial fjords (ice fields or glaciers in the hinterland), characterized by high inorganic sedimentation rates, with sediment discharges primarily from a single source (Syvitski and Murray, 1981; Curran et al., 2004).

Changes in sedimentation pattern in glacial fjords can have important consequences on other processes, as sedimentation influence some characteristics such as delivery of nutrients (Apollonio, 1973; Hooge and Hooge, 2002), physical-chemical and geotechnical properties of the seafloor (Sexton et al., 1992), aggregation and vertical flux of particles (influence on carbon flux), heat exchange with the atmosphere, and thickness of the photic zone (Svendsen et al., 2002).

Direct impact of suspended solids on the structure and distribution of planktonic and benthic communities has also been well documented (Görlich et al., 1987; Carney et al., 1999; Hop et al., 2002; Fetzer et al., 2002; Etherington et al., 2007).

The estuarine circulation in a glacial fjord during the melting season can be idealized as a subglacial buoyant jet which produces a buoyant wall plume rising along the glacier face, and a gravity current at the surface or mid-depth (Syvitski, 1989; Powell, 1990; Russell and Arnott, 2003; Salcedo-Castro et al., 2011). The behaviour of a buoyant jet depends on the balance between the buoyancy flux, given by the density difference between the plume ($\rho_0$) and the ambient fluid ($\rho_a$); and the momentum flux, represented by the initial jet velocity $u_0$. This
balance between buoyancy and momentum is represented by the Froude number (Syvitski, 1989; Powell, 1990; Russell and Arnott, 2003; Salcedo-Castro et al., 2011):

\[ Fr = \frac{u_0}{(g d (\rho_a - \rho_0) / \rho_0)^{1/2}}, \]  

(1)

where \( d \) is the opening size and \( g \) is the gravitational acceleration. Thus sub-glacial discharges can be buoyancy-dominated (\( Fr \sim 0 \)) or momentum-dominated (\( Fr \geq 1 \)) (Syvitski, 1989; Powell, 1990; Salcedo-Castro et al., 2011).

The character of the sedimentation in glacial fjords is determined by the estuarine circulation caused by the subglacial sediment-laden discharge, the presence of a stratified water mass, and the glacial front (Mackiewicz et al., 1984; Elverhøi et al., 1983). Gilbert (1982) showed that most of the sediment sinks from the gravity current (and is therefore deposited) within 15 to 20 km from the fjord head. Elverhøi et al. (1983) observed that about 90\% of the sediment input from Kongsvegen is deposited relatively adjacent to the ice front. Svendsen et al. (2002) found that during summer particulate inorganic matter (PIM) was \( \sim 0.34 \) kg m\(^{-3}\) at the glacier front and decreased to \(< 0.02 \) kg m\(^{-3}\), 10 km away.

Syvitski (1989) has pointed out that the presence of a suspended sediment load increases the initial momentum and velocity of a buoyant jet but a significant settling velocity of particles will produce a more rapid decaying of the jet velocity than that observed in a jet containing only dissolved matter. Thus it is expected that the suspended sediment will affect the buoyant discharges differently, depending on whether they are buoyancy or jet dominated. Studies of
sedimentation in glacial fjords have however been primarily focused on bulk sediment and so little is known about fine, cohesive, sediment transport in spite of its predominance in these systems (Syvitski, 1989; Curran et al., 2004). For instance, Zaborska et al. (2006) classified all sediments of the Kongsfjorden as mud, but the proportion of clay and the organic matter concentration in sediments increases with distance from the glacier. A similar conclusion was drawn by Trusel et al. (2010) who asserted that the smallest particle size fraction (silt-clay) was the predominant sediment in suspension 470 m away from the glacier. Transportation and deposition of fine-grained sediment and mud from the glacier to distal locations is primarily driven by gravity currents (Curran et al., 2004) that can maintain concentrations of fine sediments greater than 10 kg m$^{-3}$ in suspension (Mackiewicz et al., 1984).

Whereas suspended fine sand and coarse silt sink as single grains, the settling of finer silt and clay is affected by flocculation and the existence of aggregates (Syvitski, 1989; Curran et al., 2004). Flocculation is primarily dependent on sediment concentration (Mehta, 1986; Dyer, 1995; Hill et al., 1998, 2000; Shi and Zhou, 2004; Liu, 2005), but it is also influenced to a lesser extent by salinity, turbulence and other factors (Winterwerp, 2002; Dyer et al., 2002).

Field and laboratory studies of sedimentation from buoyant jets and plumes have been mainly focused on non-cohesive sediments (Carey et al., 1988; Sparks et al., 1991; Bursik, 1995; Ernst et al., 1996; Lane-Serff and Moran, 2005). Recently, Lane-Serff (2011) modeled the deposition of cohesive sediment from buoyant jets and found that the fall-speed decreases as the sediment load de-
creases. Lane-Serff also observed that the deposition rate was lower near the source but higher further away as more sediment remained in the current for longer distances.

Another process that has recently been shown to influence the sediment transport associated with buoyant plumes is convective sedimentation (McCool and Parsons, 2004). This is produced when the stratification hinders the descent speed of the sediment and, as a result, sediment concentrates along the pycnocline, until the region becomes gravitationally unstable and the inhomogeneities in the density field turn into convective cells (McCool and Parsons, 2004; Hoyal et al., 1999; Parsons and Garcia, 2001). Laboratory observations by Green (1987) about this “sediment fingering” showed that this process can be important especially in conditions of high sediment concentration, small particules and weak stratification. Parsons et al. (2001) stated that this convection occured even at sediment concentrations as low as 1 kg m\(^{-3}\), and one consequence of the convective instability of the original hypopycnal plume was the generation of a bottom turbidity current, or hyperpycnal plume that moved at moderate speeds over the bottom.

There have been some modelling efforts to study the sedimentation process in glacial fjords. Mugford and Dowdeswell (2007) used a stratigraphic simulation model that could link the environmental and climatic conditions to the geological formation of distinctive glacimarine deposits in Kangerdlugssuaq Fjord (Greenland) and McBride Glacier (Alaska). More recently, Mugford and Dowdeswell (2011) used a jet model and could reproduce some important features of the sed-
Here we carry out a fundamental numerical study of fine sediment transport associated with buoyant discharges in glacial fjords, considering a range from buoyancy to momentum-dominated conditions. We hope to capture some basic understanding about the sediment transport in glacial fjords, using a simplified configuration that does not include ambient stratification, ocean currents, or ice processes.
2. Methods

2.1. Sediment transport

In glacial fjords the freshwater source is a buoyant jet that usually enters the fjord at the base of the glacier, as subglacial discharges. The resulting vertical plume that flows along the glacier face has a typical horizontal length scale $L \sim 1$ m, that is much smaller than the vertical scale of the plume which is roughly the fjord depth, i.e. $H \sim 100$ m. The freshwater forcing in glacial fjords is, therefore, highly nonhydrostatic because $H/L \gg 1$ (Marshall et al., 1997).

Most models used in oceanography consider the hydrostatic assumption which is justified when horizontal length scales $L$ of the motion are several orders of magnitude larger than vertical length scales $H$ (Cushman-Roisin, 1994). Hydrostatic models, however, are not suitable to simulate highly nonhydrostatic processes such as convection and high-frequency gravity waves (Marshall et al., 1997), shelf/slope convection, and buoyantly driven coastal jets (Gallacher et al., 2001; Shaw and Chao, 2006). Consequently, we used a nonhydrostatic model developed by Bourgault and Kelley (2004). The model used is a two dimensional, laterally averaged model and uses a finite difference scheme with a variable mesh $z$-coordinate C-grid. The model details and experimental configuration used here are described in Bourgault and Kelley (2004) and Salcedo-Castro et al. (2011), respectively.

The numerical experiments were set in a two-dimensional configuration $(x, z)$ representing a longitudinal section of a glacial fjord, and with freshwater forcing at the glacier face. The glacier was represented as a vertical wall with a no-
slip boundary condition and the only forcing was a steady flow produced at the bottom open cells set through the glacier face. The total length of the numerical domain was 206 km and the total depth was $H = 100$ m. The numerical grid has a constant vertical resolution of $\Delta z = 1$ m. In the horizontal, the grid has a resolution of $\Delta x = 1$ m for $0 < x < 100$ m (i.e. the region of interest). For $x > 100$ m the grid size increases linearly to a maximum of $\Delta x = 5000$ m. The domain was made very long compared to the plume width such that the seaward boundary condition did not influence the results (Fig. 1).

The initial condition was set as still, uniform density ambient water and all simulations were run with a free surface and reached steady state in the region $x < 100$ m before the freshwater front reached the seaward boundary.

The module for sediment transport in the model includes an equation for the advection-diffusion of sediment concentration,

$$\frac{\partial C}{\partial t} + u \frac{\partial C}{\partial x} + (w + w_s) \frac{\partial C}{\partial z} = \frac{\partial}{\partial x} \left( \kappa_e \frac{\partial C}{\partial x} \right) + \frac{\partial}{\partial z} \left( \kappa_e \frac{\partial C}{\partial z} \right),$$

(2)

where, $C(x, z, t)$ is the sediment concentration, $\kappa_e(x, z, t)$ is the coefficients of eddy diffusivity; and $w_s$ is the sediment settling velocity.

The following expression is included to account for the modification of the equation of state for density by the presence of sediments (Wang et al., 2005):

$$\rho = \rho_w + \left( 1 - \frac{\rho_w}{\rho_s} \right) C,$$

(3)

where $\rho_w$ is the density of water and $\rho_s$ is the density of sediment. Also, the model includes the following bottom boundary condition to represent the pro-
cesses of resuspension and deposition (Partheniades, 1965; Kuijper et al., 1989; Markofsky and Westrich, 2007):

\[
\kappa_e \frac{\partial C}{\partial z} - w_s C = E_b, \quad (4)
\]

where:

\[
E_b = \begin{cases} 
E_0 \left( \frac{|\tau_b|}{\tau_c} - 1 \right) & \text{if } |\tau_b| > \tau_c \text{ (resuspension)} \\
C_b w_s \left( 1 - \frac{|\tau_b|}{\tau_c} \right) & \text{if } |\tau_b| < \tau_c \text{ (deposition)}
\end{cases} \quad (5)
\]

Here, \(E_b\) is the bottom sediment flux, \(E_0\) is the erosion coefficient, \(C_b\) is the sediment concentration at the bottom layer; and \(\tau_c\) is the critical stress for resuspension and deposition (McAnally and Mehta, 2001; van Rijn, 2007). The choice of parameters used here is shown in Table 1.

2.2. Flocculation

All runs considered the sediment grain fraction that predominates in glacial fjords, which is in the range of the silt-clay fraction (mud) (Table 2). Thus we chose a cohesive sediment whose representative particle settling velocity was roughly \(1.0 \times 10^{-5} \text{ m s}^{-1}\) (very fine silt-coarse clay with grain density of \(\sim 2650 \text{ kg m}^{-3}\)). Therefore, it was necessary to represent the process of flocculation in the model.

To represent flocculation, we used the well-known power law relationship between sediment concentration and settling velocity (Mehta, 1986) (eq. 6b), modified to account for reduced settling velocity caused by high sediment con-
centrations (Richardson and Zaki, 1954; Mehta, 1986; Puls et al., 1988) (eq. 6c):

\[
 w_s = \begin{cases} 
 w_0 & \text{if } C \leq 8.6 \times 10^{-3} \text{ kg m}^{-3} \\
 k_1 C^n & \text{if } 8.6 \times 10^{-3} < C \leq 1.7 \text{ kg m}^{-3} \\
 w_{so} (1 - k_2 C)^\beta & \text{if } C > 1.7 \text{ kg m}^{-3} 
\end{cases} 
\] (6a) (6b) (6c)

The setting for flocculation is shown in Table 3. The parameters set in Table 3 result in a maximal settling velocity of \(2.4 \times 10^{-3} \text{ m s}^{-1}\), which is in the range observed in the field (Hill et al., 1998; Shi and Zhou, 2004). The dependence of settling velocity on sediment concentration is linear up to a concentration of 1 kg m\(^{-3}\) (Fig. 2).

Experiments covering a range from buoyancy to jet dominated conditions were run. These experiments encompassed a range of Fr between 0.01 and 3.2 and are summarized in Fig. 3. Four sediment concentrations were set: 0.1, 0.1, 1, and 10 kg m\(^{-3}\). The upper end of this range of concentrations was set according to observations made by some authors (Gilbert, 1983; Mackiewicz et al., 1984; Gilbert et al., 2002).
3. Results

3.1. Plume sediment concentration

All experiments exhibited similar flow patterns: a buoyant jet issuing horizontally from the tunnel opening, a vertical plume rising attached to the “wall” that produced a lifting of the sea surface when reaching the surface and a gravity surface current that set an estuarine circulation (Fig. 4). There is a clear difference between momentum-dominated and buoyancy-dominated runs. The momentum-dominated runs showed a jet spreading horizontally on the bottom for a distance of some meters until a balance is reached as momentum is lost and buoyancy forces the jet to veer up and back to the wall to rise as a vertical plume. On the other hand, the buoyancy-dominated runs went up immediately after leaving the tunnel opening.

Flocculation did not produce any noticeable deviation of the description above for concentrations lower than 1 kg m\(^{-3}\). When the initial jet sediment concentration was 10 kg m\(^{-3}\), the experiments exhibited a different pattern. After apparently having reached steady state, some sediment commenced to settle through the water column in the far field (between 1500 and 5000 meters away of the glacier) (Fig. 5), in the form of finger-like extensions that came off the surface layer. This convective transport was preceded by subsurface higher sediment concentrations, between 0.3 - 0.4 kg m\(^{-3}\), and reached velocities higher than \(1.0 \times 10^{-2}\) m s\(^{-1}\) and involved deposition rates between \(5.0 \times 10^{-4}\) and \(1.0 \times 10^{-3}\) kg m\(^{-2}\)s\(^{-1}\). In contrast, deposition rates between 1.0 and \(8.0 \times 10^{-4}\) kg m\(^{-2}\)s\(^{-1}\) were observed above to the bottom. As sediment settled through
the water column, it was carried back to the glacier by the landward lower estuarine current and re-entrained into the vertical and horizontal plumes (Fig. 6). This process was also observed in run #09, with an initial jet sediment concentration of 1 kg m$^{-3}$.

Vertical profiles of sediment concentration for the runs with an initial jet sediment concentration of 10 kg m$^{-3}$ were obtained at a distance 10 $d$ away of the glacier (Fig. 9). The sediment concentration at the surface is higher from run10 to run 13 (increasing buoyancy-dominance) and this causes a progressive weakening of the gradient at the interface. A decrease of the sediment concentration through the water column from run10 to run 13 is seen too. It is also possible to observe the lutocline above the bottom.

Density was affected by the presence of sediment in the far field and this can be seen in Fig. 7. The sediment concentration increased and formed a thin layer of higher concentration at the base of the horizontal buoyant plume. After some time, this thin layer collapsed and sediment settled through the water column, driven by convective mixing, and forming clouds of sediment that are transported back to the glacier. This effect of sediment on the density was also observed in the near field (Fig. 8).

The maximum sediment concentration of the surface plume through a vertical cross-section at distance 10$d$ from the glacier was extracted and compared to the input concentration. The same analysis was done for a horizontal cross-section taken through the vertically rising plume at a distance 10$d$ from the bottom. The sediment concentration, nondimensionalized with the initial jet sed-
iment concentration, is lower as we move from momentum dominated (high Fr) to buoyancy dominated (low Fr) conditions in the vertical (Fig. 10(a)) and horizontal plumes (Fig. 10(b)). Low sediment concentrations at the discharge ($10^{-2} \text{ to } 10^{-1} \text{ kg m}^{-3}$) primarily affected the momentum-dominated experiments as they showed a rapid increase as response to these low concentrations. On the other hand, when the discharge carried higher sediment concentrations ($1 \text{ to } 10 \text{ kg m}^{-3}$) it was possible to observe an increasing trend in buoyancy-dominated runs, especially at higher concentrations. This response, however, is less intense as buoyancy becomes relatively more important (decreasing Fr).

The response to increasing sediment concentrations was also assessed in terms of the upward sediment transport by the vertical plume. The sediment transport was computed through the following equation:

$$F = \int_{x=0}^{x=P_{\text{edge}}} C w \, dx,$$

(7)

where $P_{\text{edge}}$ was defined as the seaward limit where $w = 0.01 w_{\text{max}}$.

The computed sediment fluxes increase as buoyancy increases (from run08 to run13)(Fig. 11). Runs 08 and 09 (momentum-dominated) showed a slight increase of sediment transport in spite of their more accentuated velocity drop because they kept relatively high sediment concentrations with respect to the sediment concentration at the initial jet coming out from the tunnel opening.

3.2. Plume velocity

Momentum-dominated runs were more sensitive to the presence of sediments (Figs. 12). Under the effect of the sediment concentration, the vertical plume ve-
locity rapidly decays in momentum-dominated conditions. In buoyancy-dominated experiments however this deceleration is noticeable only at high sediment concentrations (Figs. 12(a)). A similar pattern was observed in the maximum velocity of the surface plume, at a distance equivalent to 10 $d$ away from the glacier (Figs. 12(b)).

3.3. Plume dilution

A dilution factor (Anwar, 1973; Lee and Lee, 1998; Chen and Rodi, 1980; Huai et al., 2010) was computed to evaluate the degree of mixing along the vertical and horizontal plumes, which was defined as:

$$S = \frac{\rho_p - \rho_0}{\rho_a - \rho_p}, \quad (8)$$

where $\rho_p$ is the plume density, at a distance equivalent to 10 $d$ above the tunnel opening for the vertical plume and 10 $d$ away from the glacier for the surface gravity plume.

Similar to the case of velocities the experiments showed an increasing plume dilution as buoyancy becomes more important (decreasing Fr) (Figs. 13). The vertical (Figs. 13(a)) and horizontal (Figs. 13(b)) plume dilution was relatively unaffected by low sediment concentrations, with exception of the momentum-dominated runs (run08 and run09). As the jet sediment concentration increases, the buoyancy-dominated experiments showed a decrease in their dilution capacity. This reduction in dilution, however, is lower as the experiments are in the extreme of buoyancy-dominance.
4. Discussion

The addition of sediment produces a decrease in buoyancy and, consequently, a higher Fr number. This is relevant in glacial fjords because, as Salcedo-Castro et al. (2011) pointed out, the estuarine circulation is primarily driven by the plume buoyancy, with the plume momentum playing a secondary role. As observed in the variations of velocity, sediment concentration and dilution, however, buoyancy still remains as the main factor controlling the fine sediment transport and sediment produces significant changes only at relatively high concentrations.

The experiments showed that fine sediment can be transported a relatively long distance away of the glacier by the horizontal buoyant plume and the sediment concentration is progressively diluted by entrainment before starting to settle through the water column. Similarly, Lane-Serff (2011) observed a lower deposition rate of cohesive sediment near the origin (compared to non-cohesive sediment) but became higher further away from the source (as more sediment remains in the current for longer distances).

All experiments with jet sediment concentration of 10 kg m$^{-3}$ exhibited higher subsurface sediment concentrations at the base of the horizontal plume in the far field. This higher sediment concentration led to an instability and finally a convective transport of sediment downward through the water column. This description seems to agree well with the explanation provided by Carey et al. (1988) who asserted that the downward flux of sediment through the water column could be caused by the re-entrainment of sedimenting particles in the fluid
around the plume that increase the particle concentration of the plume margins so that it would have a density greater than either the ambient fluid or the plume interior.

The convective transport of sediment down through the water column observed in our experiments had higher vertical velocities than those caused by flocculation. This pattern was similar to the description given by McCool and Parsons (2004), who observed convective plumes that dominated sedimentation and had vertical velocities of 1-2 cm s\(^{-1}\), two orders of magnitude larger than those predicted by Stokes settling of the constituent particles. Also, surface plume concentrations as low as 380 mg L\(^{-1}\) (0.38 kg m\(^{-3}\)) were documented to support robust mixing-induced convective sedimentation (McCool and Parsons, 2004), which is in the same range observed in this study.

The process of sediment being carried back and re-entrained into the vertical and horizontal plumes has been described for non-cohesive sediments by other investigators. In plumes with concentrations greater than 10 g L\(^{-1}\), Carey et al. (1988) observed the generation of dilute downward moving flows along the side of the vertical plume. Sparks et al. (1991) described an outer region where sediment falls out from the base of a horizontal turbulent gravity current and is drawn back towards the plume by a net inflow caused by the entrainment of ambient fluid as the plume rises. Ernst et al. (1996) also observed that the reentrainment was most vigorous in runs with relatively fine-grained particles and buoyant plumes or strong jets. More recently, Cuthbertson and Davies (2008) described a tendency of settling of non-cohesive particles to be drawn
back towards the margins of the rising buoyant jet and this return flow could be sufficiently strong to re-entrain depositing particles into the rising buoyant jet. Besides, Cuthbertson et al. (2008) defined a critical distance within which particles would be reentrained back into the rising buoyant jet, whereas those settling beyond this distance will deposit to the bed. Our results, however, showed a combination of these processes where sediment is transported from the far field back to the vertical plumes and, at the same time, part of the sediment is being deposited on the bed.

The experiments with initial sediment concentrations of 10 kg m\(^{-3}\) in the issuing jet had sediment concentrations at the surface plume between 0.7 - 1 kg m\(^{-3}\) which yield settling floc velocities of 1 - 1.4 mm s\(^{-1}\). This range is similar to what has been observed in fjords and other estuaries by some authors. Hill et al. (1998) found that the predicted settling velocity of a 1 mm diameter floc is 1.5 mm s\(^{-1}\). Shi and Zhou (2004) calculated settling velocities from 0.4 to 4.1 mm s\(^{-1}\) for the point-sampled data set, and from 1.0 to 3.0 mm s\(^{-1}\) for an acoustically measured data set.

Despite the good representation of fine sediment transport by the model, there are other processes not included and that could predominate during certain stages and in some regions of the jet and vertical and horizontal plumes. Verney et al. (2009) demonstrated that turbulent intensity is one of the main determining factors of maximum floc size. In this sense, Pejrup and Mikkelsen (2010) has shown that the with the inclusion of turbulence, an improvement of up to 72% has been found in explaining the variation in settling velocity. In this sense, Domack et al.
(1994) stated that turbulent mixing near the seafloor can play an important role in the transport and break-up of floccules.

In our simulations, the background environment was considered motionless, without wind or tides that produce background turbulence. This is justified for high latitude systems where the tidal range is narrow when compared to other estuaries. In this sense, wave-associated resuspension is not considered important either, as we represented a glacial fjord adjacent to a tidewater outlet glacier where shallow areas and tidal flats are practically inexistent. Further studies should consider the inclusion of turbulence and mixing associated with waves, which is expected to produce somewhat different results. Simulations with realistic tidal forcing and stratified conditions are left to further studies.
5. Conclusion

Momentum-dominated conditions are more sensitive than buoyancy-dominated conditions to the presence of sediment in the buoyant jet discharging into the ambient water. Therefore this type of experiments shows response at even low sediment concentrations. On the other hand, buoyancy-dominated experiments exhibited noticeable changes only at high sediment concentrations and this response was less intense as buoyancy increased (Fr becoming smaller).

Cohesive sediments do not settle in the near field but it is transported to the far field and settle there. Then it is carried back to the glacier and reentrained into the vertical and horizontal plumes.

The density field is affected by the presence of sediment, as instabilities were produced by higher subsurface sediment concentrations observed at the interface between the upper and lower layer, and clouds of this denser water (and sediment) go down convectively through the water column.

Convective sedimentation proved to be a more efficient mechanism of vertical sediment transport of fine sediment, compared to individual particles settling and flocculation.
6. Acknowledgements

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7. References


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Sediment concentration profiles (red) taken at a distance equivalent to 10 \(d\), where \(d\) is the opening diameter (Initial jet sediment concentration: 10 kg m\(^{-3}\)). Experiments without flocculation are included (black lines) for comparison.

Effect of sediment concentration on the vertical (a) and horizontal (b) plume concentration for different Fr numbers at a distance equivalent to 10 \(d\), where \(d\) is the opening diameter.

Vertical sediment flux (nondimensionalized with the initial jet sediment flux) computed at 10 \(d\), where \(d\) is the opening diameter.

Effect of sediment concentration on the vertical (a) and horizontal (b) plume velocity for different Fr numbers at a distance equivalent to 10 \(d\), where \(d\) is the opening diameter.

Effect of sediment concentration on the vertical (a) and horizontal (b) plume dilution for different Fr numbers at a distance equivalent to 10 \(d\), where \(d\) is the opening diameter.
Figure 1: Schematic representation of a glacial fjord, showing parameters considered in numerical experiments.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\rho_s$ (kg m$^{-3}$)</td>
<td>2,650</td>
</tr>
<tr>
<td>$\rho_w$ (kg m$^{-3}$)</td>
<td>1,000</td>
</tr>
<tr>
<td>$w_0$ (m s$^{-1}$)</td>
<td>0.00001</td>
</tr>
<tr>
<td>$E_0$ (kg m$^{-2}$ s$^{-1}$)</td>
<td>0.0001</td>
</tr>
<tr>
<td>$\tau_c$ (Pa)</td>
<td>0.3</td>
</tr>
</tbody>
</table>
Table 2: Typical values of sediment concentration and grain size found in glacial fjords

<table>
<thead>
<tr>
<th>Location</th>
<th>Reference</th>
<th>Concentration range (kg m$^{-3}$)</th>
<th>Size range</th>
</tr>
</thead>
<tbody>
<tr>
<td>Arthur Harbor, Antarctica</td>
<td>Ashley and Smith (2000)</td>
<td>0.003 - 0.035</td>
<td>clay/silt (30:60)</td>
</tr>
<tr>
<td>MacBride Glacier, Alaska</td>
<td>Cowan and Powell (1990)</td>
<td>0.45 - 0.50</td>
<td></td>
</tr>
<tr>
<td>MacBride Glacier, Alaska</td>
<td>Cowan et al. (1988)</td>
<td>0.5 - 2</td>
<td>99.6% &lt; 63µm</td>
</tr>
<tr>
<td>Hubbard Glacier</td>
<td>Curran et al. (2004)</td>
<td>0.01 - 0.035</td>
<td></td>
</tr>
<tr>
<td>Cierva, Brialmont, Lester Cove, Antarctica</td>
<td>Domack and Williams (1990)</td>
<td>0.0006 - 0.008</td>
<td></td>
</tr>
<tr>
<td>Brialmont Cove, Antarctica</td>
<td>Domack et al. (1994)</td>
<td>0.00075 - 0.0041</td>
<td>2 – 10µm</td>
</tr>
<tr>
<td>Watts Glacier and Coronation-Maktak fjords</td>
<td>Dowdeswell (1986)</td>
<td>&lt; 4µm</td>
<td></td>
</tr>
<tr>
<td>Spitsbergen</td>
<td>Elverhøi et al. (1983)</td>
<td>0.02 - 0.5</td>
<td></td>
</tr>
<tr>
<td>Kongsfjorden, Spitsbergen, Norway</td>
<td>Fetzer et al. (2002)</td>
<td>90% clay/silt (50:50)</td>
<td></td>
</tr>
<tr>
<td>Coronation Fjord, Baffin Island</td>
<td>Gilbert (1982)</td>
<td>30 – 100µm</td>
<td></td>
</tr>
<tr>
<td>Pangnirtung Fjord</td>
<td>Gilbert (1978)</td>
<td>65 – 90% clay/silt</td>
<td></td>
</tr>
<tr>
<td>Hornsund Fjord, Spitsbergen</td>
<td>Görlich et al. (1987)</td>
<td>0.01 - 1</td>
<td></td>
</tr>
<tr>
<td>Itirbilung Fjord, Baffin Island</td>
<td>Hein and Syvitski (1992)</td>
<td>0.001 - 0.028</td>
<td>sand/mud (50:50)</td>
</tr>
<tr>
<td>Blue Fjord, Alaska</td>
<td>Hoskin et al. (1978)</td>
<td>0.01 - 0.3</td>
<td>46 – 53µm</td>
</tr>
<tr>
<td>Muir Inlet, Alaska</td>
<td>Mackiewicz et al. (1984)</td>
<td>65 – 90% 4 – 16µm</td>
<td></td>
</tr>
<tr>
<td>Nordaustlandet tidewater ice cap, Svalbard</td>
<td>Pfriman and Solheim (1989)</td>
<td>0.001 - 0.028</td>
<td></td>
</tr>
<tr>
<td>Lange Glacier, Antarctica</td>
<td>Pichlmaier et al. (2004)</td>
<td>0.007 - 0.011</td>
<td></td>
</tr>
<tr>
<td>Martel Inlet, Antarctica</td>
<td>Pichlmaier et al. (2004)</td>
<td>0.01 - 0.015</td>
<td></td>
</tr>
<tr>
<td>Kongsfjorden, Svalbard, Norway</td>
<td>Svendsen et al. (2002)</td>
<td>&lt;0.02 - 0.34</td>
<td></td>
</tr>
<tr>
<td>Coronation Fjord, Baffin Island</td>
<td>Syvitski (1989)</td>
<td>0.01 - &gt;0.120</td>
<td></td>
</tr>
<tr>
<td>Billefjorden, Svalbard</td>
<td>Szczuciński et al. (2009)</td>
<td>&gt; 90% clay/silt (50:50)</td>
<td></td>
</tr>
<tr>
<td>Kongsfjorden, Svalbard</td>
<td>Trusel et al. (2010)</td>
<td>0.008 - 0.157</td>
<td></td>
</tr>
<tr>
<td>Kongsfjorden, Svalbard</td>
<td>Zaborska et al. (2006)</td>
<td>&gt; 0.3</td>
<td>&gt; 90% clay/silt (10:90)</td>
</tr>
<tr>
<td>Kongsfjorden, Svalbard</td>
<td>Zajaczkowski (2008)</td>
<td>0.35 - 0.46</td>
<td></td>
</tr>
</tbody>
</table>
Table 3: Parameters used for flocculation in the model

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$k_1$</td>
<td>0.14</td>
</tr>
<tr>
<td>$n$</td>
<td>1.04</td>
</tr>
<tr>
<td>$k_2$</td>
<td>0.0001</td>
</tr>
<tr>
<td>$w_{so}$ (m s$^{-1}$)</td>
<td>0.0026</td>
</tr>
<tr>
<td>$\beta$</td>
<td>4.65</td>
</tr>
</tbody>
</table>

Figure 2: Settling velocity as function of sediment concentration.
Figure 3: Effect of sediment concentration on Fr number.
Figure 4: Typical sequence of sediment concentration in a momentum-dominated jet issuing into the ambient denser water (run08, initial jet sediment concentration: 0.1 kg m$^{-3}$).
Figure 5: Sequence of sediment concentration in the gravity plume spreading at the surface and settling of sediment in the far field (run10, initial jet sediment concentration: 10 kg m$^{-3}$).
Figure 6: Sequence of sediment concentration in the gravity plume spreading at the surface and settling of sediment in the near field (run10, initial jet sediment concentration: 10 kg m$^{-3}$).
Figure 7: Sequence of density anomaly field and changes associated with settling of sediment in the far field (run10, initial jet sediment concentration: 10 kg m$^{-3}$).
Figure 8: Sequence of density anomaly field and changes associated with settling of sediment in the near field (run10, initial jet sediment concentration: 10 kg m\(^{-3}\)).
Figure 9: Sediment concentration profiles (red) taken at a distance equivalent to 10 \( d \), where \( d \) is the opening diameter (Initial jet sediment concentration: 10 kg m\(^{-3}\)). Experiments without flocculation are included (black lines) for comparison.
Figure 10: Effect of sediment concentration on the vertical (a) and horizontal (b) plume concentration for different Fr numbers at a distance equivalent to $10d$, where $d$ is the opening diameter.

Figure 11: Vertical sediment flux (nondimensionalized with the initial jet sediment flux) computed at $10d$, where $d$ is the opening diameter.
Figure 12: Effect of sediment concentration on the vertical (a) and horizontal (b) plume velocity for different Fr numbers at a distance equivalent to $10 \, d$, where $d$ is the opening diameter.

Figure 13: Effect of sediment concentration on the vertical (a) and horizontal (b) plume dilution for different Fr numbers at a distance equivalent to $10 \, d$, where $d$ is the opening diameter.