# Advective sorting of silt by currents: a laboratory study

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**Abstract**

Accumulations of fine sediments along continental shelf and deep-sea bathymetric contours, known as contourite drifts, form a sedimentary record that is dependent on oceanographic processes such as ocean-basin-scale circulation. A tool used to aid in interpretation of such deposits is the sortable silt hypothesis, which suggests that the mean size of the sortable silt (silt from 10-63 $\mu$m) within a deposit can be used as a proxy for current velocity. While the hypothesis has been applied to numerous drift deposits, it has not been extensively explored. Slow deposition rates of contourite drift systems make it difficult to study in the deep ocean, and past laboratory studies have not tested the full range of conditions or mechanisms that could lead to sorting. This study uses flume experiments and theory to examine how the mean sortable silt in a deposit is related to current velocity under the action of advective depositional sorting. Tests were conducted for a fixed amount of time with four suspended sediment mixtures and current velocities typical of deep-sea settings (5-25 cm/s). Developed beds were sampled at fixed locations from the entrance and sized. The deposit grain size fined downstream and coarsened with increasing velocity at a particular distance from the inlet. Simple theory was able to capture the observations. Regardless of bed morphology or source sediment mixture, the mean sortable silt in the deposit was related to velocity at a particular flume location across all sediment mixtures. The slope of the relationship between velocity and size was dependent on the distance between the inlet and location of interest. Despite the simplified nature of the study, and the limitations regarding

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the presumed variability in natural systems, these findings broadly support the validity of mean sortable as a proxy for paleocurrent velocity at a distance along a depositing current.

*Keywords:*
advective sorting, clay, contourite drift, silt, sortable silt

## 1. Introduction

Sedimentary deposits composed of particles $< 63\ \mu$m in size (i.e., mud/mudstone) make up the majority ($> 60\%$) of Earth's sedimentary record (Schieber, 1998), contain valuable archives of past environments and climate conditions (e.g., Knutz, 2008), and host important energy resources (e.g., Jarvie et al., 2007; Slatt, 2011). However, the physical processes that erode, transport, and deposit fine-grained sediment are still poorly understood compared to other sedimentary deposits (Schieber, 2011). Significant progress has been made over recent years demonstrating that clay ($< 4\ \mu$m) can behave hydrodynamically similar to sand-size grains (e.g., bed load transport and associated bedform development) as a result of aggregation into larger, composite particles (up to 0.1-1.0 mm in size; Schieber et al., 2007); however, there has been comparably less focus on the silt fraction (4-63 $\mu$m). The 'sortable silt' proxy, which was developed by the palaeoceanography community to aid reconstruction of past deep-ocean current activity (McCave et al., 1995; McCave and Hall, 2006), provides a framework to further investigate fundamental questions pertaining to the way in which transport and deposition of silt could lead to sorting within a deposit under a specific set of conditions.

Currents in marine environments are responsible for the erosion, transport, and deposition of fine-grained sediment. In some areas, including coastal zones, continental shelves and shallow seas, and the deep ocean, significant accumulations of muddy deposits develop over geological timescales ($> 10^6$ yr) and, thus, represent archives of current history. For example, in the deep ocean, long-lived ocean-basin-scale boundary currents are a key component of global ocean circulation (Broecker et al., 1998) and have been linked to contourite 'drift' deposits that exceed a kilometre in thickness and extend for hundreds of kilometres (Heezen et al., 1966; Rebesco et al., 2014). Paleoceanographers have been especially interested in reconstructing bottom-current history from contourite drift deposits because of the linkage of abyssal currents to thermohaline circulation and global climate. To aid in reconstruction, McCave et al. (1995) proposed a sedimentological proxy for changes in current velocity that relate the size of non-cohesive silt to the speed of the eroding or depositing flow. Specifically, McCave et al. (1995) hypothesized that particles within the size range of 10 to 63 $\mu$m, defined as 'sortable silt' (SS), sort by size with current velocity during selective deposition and selective erosion (winnowing) (see McCave and Hall, 2006, for a comprehensive review). Key statistics associated with this size range that have been linked to velocity of ocean bottom currents include the mean SS, $SS_{mean}$, and the SS percentage, $SS\%$ (McCave and Andrews, 2019).

Despite the widespread application of the SS hypothesis to deep-sea sediment core samples of mostly Quaternary deposits (e.g., Kleiven et al., 2011; Thornalley et al., 2013), and some attempts at field calibration (McCave et al., 2017), the proxy remains largely untested under controlled conditions. Slow deposition rates of contourites (2-10 cm/kyr) make it difficult to robustly explore the hypothesis in the deep ocean, and the limited number of laboratory studies that have been reported have not explored the full range of possible conditions or mechanisms that could lead to sorting in the deep sea. In addition, the general lack of knowledge regarding sediment transport and deposition mechanisms in these systems highlights the need for fundamental experiments that isolate specific processes. Only two known studies have sought to explicitly examine ideas related to the SS hypothesis. The first, Law et al. (2008), set out to assess whether or not silt and clay beds are subject to sorting through the process of selective erosion of fine grains. Using core samples from the Gulf of Lions and a fluid shearing device known as a Gust chamber, Law et al. (2008) found that sediment samples with less than 7.5% clay were subject to erosional sorting via selective erosion (or winnowing), but that beds with more than 7.5% clay were not susceptible to sorting. In cases with clay content greater than 7.5%, the core surface tended to fail en masse rather than grain by grain. The study highlighted the possibility of sorting by winnowing in non-cohesive beds and the importance of clay (present and abundant in the majority of deep-sea drifts) in modulating erosional sorting. However, the study did not examine sorting via selective deposition.

Adding to the work of Law et al. (2008), Hamm and Dade (2013) used controlled laboratory experiments conducted in a recirculating, oval-track flume to examine sorting by a unidirectional current. Rather than focusing on selective erosion only, those authors examined the use of SS as a proxy over current velocities that allowed for both deposition and re-entrainment of previously deposited material in a current. In the experiment, silt-sized glass spheres were used (diameters between 13-44 $\mu$m) for sediment, with no clay-sized sediment added, and current velocities between 20 and 53 cm/s. For these conditions, Hamm and Dade (2013) reported no evidence of significant sorting of grain size within the bed as a function of current velocity. While these results did not confirm the SS hypothesis originally proposed by McCave et al. (1995), Hamm and Dade (2013) suggested that the study was not able to make any conclusions about the appropriateness of the SS proxy under the conditions of advective down-current sorting of silt from a sediment source due to the recirculating nature of the flume.

Both the Law et al. (2008) and Hamm and Dade (2013) studies have focused on conditions that produce re-entrainment or erosion of bed material. Yet, it is likely that the majority of deep-sea drift strata develops under largely depositional conditions (current velocity $<$ 25 cm/s). The field studies of Ledbetter (1986) and McCave et al. (2017) suggest that size statistics in the silt range may indeed serve as a viable proxy for oceanic bottom current velocity. Yet the velocity range over which their data suggests that there is a correlation between velocity and silt size statistics is for $U <$ 25 cm/s; a velocity range that has not been tested experimentally. Furthermore, if the sampled deposits develop under down-current dispersal of suspended sediment that originated from localized and short-lived erosional events (Richardson et al., 1993), then it is possible that down-current advective transport from a source may in fact be an important mechanisms of sorting (a mechanism not examined in either Law et al. (2008) or Hamm and Dade (2013)).

The SS hypothesis stands as a valuable potential proxy for bottom current speed. The laboratory study presented here focuses exclusively on one potential mechanism of sorting. The mechanism we focus on is downstream advective sorting from a source of suspended sediment under largely depositional conditions using sediment composed of different mixtures of silt and clay. As such, a different mechanism of sorting is examined under lower velocities than has been tested in past experiments. Specific aims of this study are to examine whether or not the mean of the sortable silt within a deposit is related to average current velocity, and to examine how any relationship that does exist changes as a function of distance from the sediment input location. These aims are met using laboratory flume experiments and simple theory. Both show support for the use of SS as a proxy for current velocity in systems dominated by advection but does not take into account sorting via winnowing. Thus, as part of this work, potential limitations of these findings are also discussed with respect to deposits from natural systems.

## 2. Methods

### *2.1. Approach*

This study seeks to investigate, via a flume study: (i) the functionality between the mean SS, $SS_{mean}$, and current velocity, $U$, at a particular location; and (ii) how that functionality changes with the distance from the sediment input location and the initial composition of that sediment. The current velocities we are interested in correspond to those primarily observed in oceanic bottom currents, i.e., $U = 0$ to 25 cm/s (McCave et al., 2017). Velocities in this range are sufficient to move silt-sized material, but they are generally insufficient to cause large scale erosion or pull material up into suspension (McCave et al., 1995; Niño et al., 2003; Hamm et al., 2009; Hamm and Dade, 2013). Therefore, the experiments are net depositional with any sorting that occurs likely coming from selective deposition or re-entrainment of freshly deposited grains rather than erosion of a preexisting bed.

Laboratory studies using cohesive sediments and silts have been conducted in a range different flume types, e.g., traditional recirculating laboratory flumes (Einstein and Krone, 1962), annular flumes (Haralampides et al., 2003; Partheniades, 2006; Lau and Krishnappan, 1992), enclosed shear chambers (Teeter, 1997; Law et al., 2008), racetrack flumes (Schieber et al., 2007; Hamm et al., 2009; Hamm and Dade, 2013; Yawar and Schieber, 2017), and non recirculating flumes (Dixit et al., 1982). Each flume type has its own set of strengths and weaknesses when it comes to studying the movement of sediment. A traditional recirculating flume works well for sand and gravel studies, but the high shear stress encountered in the pumps makes it less ideal for studying cohesive sediment behaviour when flocculation could be an important process of consideration. In addition, deposition of fine sediment within the flume system (flume, tailbox, pipes, and headbox) makes it difficult to maintain a constant concentration at the flume inlet (Mooneyham and Strom, 2018). Annular flumes are ring-shaped channels of water that counter rotate to induce a constant shear across the bed of the flume. While an annular flume can maintain constant shear over long periods of time without passing sediment through pumps and pipes, it cannot be used to examine changes in the deposition with distanced traveled from a source because the sediment in suspension keeps getting wrapped around the same section of bed. Oval-shaped 'racetrack' flumes are similar in that no pumps or pipes are involved and material is continuously moved around a closed circuit. Such systems work well for developing equilibrium conditions between deposition and erosion, but they also do not allow for examination of how the distance from the input alters the nature of the deposit. Taking these limitations into consideration, a linear flow-through flume was chosen here in which no water or sediment is recirculated back to the inlet. While this type of flume is resource intensive, it does allow for a constant upstream concentration boundary condition and the development of spatial patterns in the deposit.

Four different sediment mixtures (pure clay, pure silt, 2:1 silt to clay and 1:1 silt to clay) were used in the experiments. In all cases, a well-mixed suspension of water and sediment was fed at a constant rate at the upstream end of the flume. Beds developed downstream with spatially varying patterns as sediment deposited from flow. The deposited bed that remained at the end of each experiment was then sampled at 1.52 m (5 ft) increments from the inlet and measured with a SediGraph 5120 particle size analyzer (Micromeritics, Norcross, GA, USA) to allow for calculation of $SS_{mean}$. Current velocities for the experiment ranged from 5 to 25 cm/s.

### 2.2. Depositional theory

This section develops simple theory to predict the size distribution of silt in a deposit and thereby an initial prediction regarding the behaviour of $SS_{mean} = SS_{mean}(x, U)$. Comparing the experimental results to the theory provides the opportunity to extend the laboratory results to larger scales if the model is found to describe the data well. The model also can be used to help inform what processes are important to consider if the experimental data deviates from the expectation set by the simplified theory.

The context of the model is that of depositional sorting in a boundary layer flow. As such, the model's foundation is provided by the one-dimensional (layer-averaged) advection-dispersion equation for suspended sediment of grain size fraction $i$:

$$\frac{\partial(AC_i)}{\partial t} + \frac{\partial(AUC_i)}{\partial x} = \frac{\partial}{\partial x}\left(AK_x\frac{\partial C_i}{\partial x}\right) + b\left(E_{b,i} - D_{b,i}\right) \tag{1}$$

where $A$ is the cross sectional flow area of the current containing sediment, $b$ is current width, $C_i$ is the layer-averaged volume concentration of suspended sediment (the total concentration is $C = \sum C_i$), $K_x$ is the dispersion coefficient, $E_{b,i}$ is an erosive flux and $D_{b,i}$ is a depositional flux for grain size fraction $i$ across the sediment-water interface.

Assuming steady, uniform hydraulics and a rectangular or top-hat current cross-sectional area with a constant sediment input, Eq. 1 simplifies to:

$$bhU\frac{\partial C_i}{\partial x} = b\left(E_{b,i} - D_{b,i}\right) \tag{2}$$

Here $h$ is the flow thickness with the cross-sectional flow area being defined as $A = bh$. The solution of Eq. 2 yields $C_i = C_i(x)$ if models for the deposition and erosion flux are specified. Typically the maximum deposition flux for grain size fraction $i$ is taken to be $D_{b,i} = \alpha_i w_{s,i} C_{b,i}$, where $w_{s,i}$ is the settling velocity for size $i$ and $C_{b,i}$ is the near-bed concentration of size fraction $i$; $\alpha_i$ can be thought of as the ratio of the true depositional velocity of size fraction $i$ divided by the still water settling velocity, $w_{s,i}$. $E_{b,i}$ is often modeled as $E_{b,i} = w_{s,i}E_{s,i}$ where $E_{s,i}$ is a dimensionless erosion or entrainment velocity.

To solve Eq. 2, constant values or closure equations are needed for $\alpha_i$ and $E_{s,i}$. When the particle diameter is less than the thickness of the viscous sublayer, the experiments of Hamm et al. (2009) suggests that $\alpha_i$ and $E_{s,i}$ are both controlled by the ratio of the viscous lift to the particle's gravitational body force. In such cases, the analytical solution for viscous lift given by Saffman (1965) for small particle Reynolds numbers yields a ratio of these two forces that is proportional to $u_*^3/(g'\nu)$; here $u_*$ is the friction velocity and $g' = gR_s$ with $R_s$ being the submerged specific gravity of the sediment ($R_s = 1.65$). Hamm et al. (2009) referred to this ratio as the Saffman parameter, $S \equiv u_*^3/(g'\nu)$. The analytical expression for the gravitational and lift forces are both dependent on particle diameter. However, both forces are proportional to $d^3$, resulting in the ratio of the forces being independent of particle size. Based on curve fitting, Hamm et al. (2009) found that their experimental data was best described with:

$$\alpha = 1 - S \tag{3}$$

$$E_s = S^{5/2} \tag{4}$$

For the simplified model used here, Eqs 3 and 4 are adopted, resulting in size-independent $\alpha_i$ and $E_{s,i}$ values. Assuming a well-mixed condition for suspended sediment over the thickness of the turbid boundary layer ($C_{b,i} = C_i$), the deposition and entrainment fluxes become:

$$D_{b,i} = \alpha w_{s,i} C_i \tag{5}$$

$$E_{b,i} = w_{s,i} E_s \tag{6}$$

Therefore, the net rate of deposition to, or accumulation in, the bed of sediment of size fraction $i$ is:

$$D_{b,i} - E_{b,i} = w_{s,i}(\alpha C_i - E_s) \tag{7}$$

Using the rate of accumulation, the total fraction of the bed material of size $i$ can be defined as:

$$f_{b,i} = \frac{w_{s,i}(\alpha C_i - E_s)}{\sum w_{s,i}(\alpha C_i - E_s)} \tag{8}$$

Values of $f_{b,i}$ as a function of distance are what one needs to develop predictions for the spatial arrangement of $SS_{mean}$ in the flume or a boundary current. $f_{b,i} = f_{b,i}(x)$ can be obtained from the solution of Eq. 2 using Eqs 5 and 6 for the exchange rates of material at the water-sediment interface. The result is:

$$C_i = \left(C_{i,0} - \frac{E_s}{\alpha}\right) e^{-\frac{\alpha w_{s,i}}{U}\left(\frac{x}{h}\right)} + \frac{E_s}{\alpha} \tag{9}$$

In Eq. 9, $x/h$ is the scaled downstream distance. Assuming consistency between $\alpha w_{s,i}/U$ in the laboratory and field, the scaled downstream distance becomes the primary ratio for mapping the calculations from the laboratory scale to the field. Alternatively, the ratio of $hU/(\alpha w_{s,i})$, or the horizontal advective length scale for sediment of size $i$, $L_i = hU/(\alpha w_{s,i})$, can also be used in Eq. 9 to scale the downstream distance:

$$C_i = \left(C_{i,0} - \frac{E_s}{\alpha}\right) e^{-\frac{x}{L_i}} + \frac{E_s}{\alpha} \tag{10}$$

When coupled with a settling velocity equation (e.g., Ferguson and Church, 2004), the model (Eqs 8 and 9 or 10) predicts the silt size distribution in the bed as a function of the distance from the input, $x$, the current velocity, $U$, the current thickness, $h$, the size distribution of the source material, and the Saffman parameter (or $u_*$, which the authors link to $U$). An implicit assumption embedded in the model is that deposited material is immobile; that is, the model does not account for bed load transport.

This simplified model was used to examine $f_{b,i} = f_{b,i}(x, U)$ and the resulting $SS_{mean} = SS_{mean}(x, U)$ over the $x$ and $U$ values expected in the experiments (Figs 1 and 2); the exact definition of $SS_{mean}$ is given in the *Calculation of the mean sortable silt* section. These calculations were made using: (i) a synthetic initial SS size distribution based on the input sediment; and (ii) the $u_* = u_*(U)$ relation from these experiments. The synthetic SS distribution reasonably mimics the average SS size distribution used in the experiments and was developed assuming that the natural log transformed grain sizes are normally distributed with a mean of $\theta_m = 3.2$ (24.5 $\mu$m) and a standard deviation, $\sigma = 0.8$ ($\theta_m - \sigma = 11$ $\mu$m and $\theta_m + \sigma = 54.5$ $\mu$m). For context, $u_*$ values from these experiments produced Saffman numbers between 0.002 to 0.15, and therefore $\alpha$ values ranging from 1 to 0.85 with $E_s \approx 0$.

The model predicts a decrease in $SS_{mean}$ with distance from the flume inlet (or source of suspended sediment), $x$, as one would expect for downstream fining under advective sorting (Fig. 1). Increases in either $h$ or $U$ result in coarsening at a given location (Eq. 9) as a result of the increase in the advective length scale, $L_i$; doubling either $h$ or $U$ produces a doubling of $L_i$. For fixed $x$ (that is, at a station) and $h$, $SS_{mean}$ increases with $U$ (Fig. 2). The model predicts that the relationship between $SS_{mean}$ and

$U$ becomes more linear with increased distance from the input (Fig. 2). In addition, the slope of the $SS_{mean}$ and $U$ relation first increases with distance and then decreases as grain sizes available for deposition become depleted (Fig. 2). The outcome of this simple analysis suggests that a linear relationship between the average silt size (Ledbetter, 1986) or $SS_{mean}$ (McCave et al., 2017) might apply best to locations farther away from the initial sources of suspended sediment if advective sorting is responsible for the building the deposit.

### 2.3. Experimental equipment and materials

All experiments were conducted in a 18 cm wide, 9.14 m long, tilting acrylic flow-through flume (Fig. 3). The inflow for the system is controlled with a constant-head tank and valve on the inflow line. Water from the constant-head tank is discharged to a mixing tank, where sediment is added via a calibrated AccuRate dry material feeder. The suspension then flows from the mixing tank to the flume headbox, through the flow straighteners, down the length of the channel, and into a settling basin before being discharged to the drain. Uniform flow is maintained over the length of the channel through adjustment of the channel slope and a series of removable vertical bars at the flume outlet. The maximum amount of flow that can be put through the system is dependent on the volume in the storage tank, the volume flow rate of water that can be added to the storage tank, and the capacity of the drain line in the laboratory. Taken together, the maximum flow velocity that can be sustained with the system is 28 cm/s at a flow depth of 3.4 cm for 2 hours, which yields a maximum functional discharge of 1.7 L/s.

Other equipment used in the experiments included an overhead camera (Canon 80D; Canon, Tokyo, Japan) attached to a sliding rail and Campbell Scientific Optical Backscatter Sensors (OBS; Campbell Scientific Limited, Loughborough, UK) 3+ probes (Fig. 3). The camera was used during the experiments to observe, when possible, the development and movement of the bed through video and time-lapse photography. The camera was also used to develop a mosaic of the entire bed after water had been drained from the flume at the completion of each run. The OBSs were installed at the up and downstream ends of the flumes for a subset of runs to monitor concentration. These were calibrated beforehand using each of the sediment mixtures over a large range of concentrations. All regressions for the OBSs were done with 18 or more points, and $R^2$ values for each exceeded 0.99.

The sediment used in the experiments included crushed silica (silt) sourced from US Silica under the name SIL-CO-SIL, kaolinite supplied by Georgia Kaolinite, and a 100% non-treated sodium bentonite of the name Aquagel Gold Seal. Four mixtures were tested at varying flow velocities: 100% silica silt, 5:4:1 silica silt:kaolinite:bentonite, 5:8:2 silica silt:kaolinite:bentonite, and 4:1 kaolinite:bentonite. These sediment mixtures will hereafter be referred to as pure silt, silt to clay 1:1, silt to clay 1:2, and pure clay respectively. These clay:silt ratios resemble the range observed in muddy and silty contourites (Rebesco et al., 2014; McCave and Hall, 2006) as well as those tested in flocculation experiments that found silt to entrain into flocs in a depositional environment (Tran and Strom, 2017); the type of bentonite used in the present study is different from that of Tran and Strom (2017). Grain size distributions of the input sediment mixtures used for each experiment can be seen in Figure 4.

Each experiment proceeded by setting the discharge to the desired rate, checking for the development of uniform flow in the channel and adjusting the number of bars at the flume exit as needed, followed by engagement of the sediment feeder. All experiments were run for a duration of 2 hours. Through preliminary experiments 2 hours was determined to be a sufficient amount of time to accumulate enough sediment in the bed for sampling. The conditions used in the experimental matrix are given below following a discussion on results from a set of preliminary tests.

Flow velocity, $U$, was calculated for each run using the measured volumetric discharge, $Q$, the measured flow depth, $h$, and known flume width, $w$, as $U = Q/(hw)$. Shear velocity, $u_*$, values for each case were obtained by solving the smooth-wall Keulegan resistance equation,

$$\frac{U}{u_*} = \frac{1}{\kappa} \ln\left(\frac{hu_*}{\nu}\right) + 5.5 - \frac{1}{\kappa} \tag{11}$$

where, $\kappa = 0.4$ is the von Karman constant and $\nu$ is the kinematic viscosity of the water. The bed shear stress, $\tau_B$ is related, by the definition of the friction velocity, as $\tau_B = \rho u_*^2$.

### 2.4. *Flocculation potential of the sediment mixtures and the role of clay hydration time and salinity in the experiments*

The flocculation behaviour of the various mixtures that included clay was tested in a separate set of flocculation mixing-tank experiments similar to those conducted by Tran and Strom (2017). In the mixing-tank experiments, the size distribution of the material in suspension is measured as a function of mixing time using a specially designed microscope system that allows for measurement of particles and flocs in suspension without removing samples from the tank (see Tran and Strom (2017) for details). The purpose of the mixing-tank experiments was to determine if the flocculation potential of the sediment was strongly influenced by: (i) the wetting time of the clay prior to being introduced to the flume; and (ii) the presence or absence of salt. The experimental parameters investigated included: sediment type (the silt and kaolinite mix with and without the bentonite), wetting time (a short wetting time and a 24 hr wetting time), and salinity (0 and 5 ppt). This set of parameters resulted in eight mixing-tank experiments. The short wetting time samples were obtained by turning on the flume, sediment feeder, and water-sediment mixing system at the rates used in the flume experiments, collecting samples of suspended sediment issuing from the flume inlet, and then transferring those samples to the flocculation mixing tank. The 24 hr hydration samples were collected in a similar way but were left to sit in the mixing tank water for 24 hours before starting the mixer to observe flocculation. For each experiment, the mixer was set as low as possible while still being able to maintain the material in suspension. Size distributions of the material in suspension were recorded every minute for two hours (sample images from the experiments can be found in the supplemental material).

For all experiments, particles and aggregates consisting of clay and silica were present in suspension. However, no significant flocculation was observed in any experiment. The size distributions of the suspensions were consistent both through time in each experiment and across all experiments. Clay in suspension did exist in a range of aggregate sizes, but these aggregates were small and more compact than the more loosely bound flocs of Tran and Strom (2017). Furthermore, no binding of the silica silt particles in the clay aggregates was observed. Samples extracted near the bed and imaged with the microscope system from two of the flume experiments showed similar sized material and a general lack of large, low-density flocs capable of binding up silt particles. Based on the mixing-tank experiments, it is concluded that the clay and silt mixtures emanating from the flume inlet would not be significantly different if the sediment mixture was pre-hydrated for more time or salt was added to the water. Knowing this is advantageous because of the difficulty involved with wet feeding a saline, uniform sized suspension of material over the course of two hours. In addition, the experiments show that the potential for silt particles to bind within clay flocs is low for these particular sediment mixtures.

### *2.5. Deposit sampling and grain-size measurement*

Bed samples were collected at the end of each 2-hr experiment after most of the water in the flume had been carefully drained from the flume. Samples were collected at five stations at distances of 1.52 m (5 ft), 3.05 m (10 ft), 4.57 m (15 ft), 6.10 m (20 ft), and 7.62 m (25 ft) from the inlet (Fig. 3). To ensure the capture of all of the fine sediment, the samples were collected using a large syringe over a 5 cm x 5 cm patch of bed (or an area large enough to sample a minimum of 1 g of sediment by dry weight). Each sample contained both sediment and water. These samples were stored in labelled vials for sizing at a later time and the syringe was flushed with clean water several times between each sample to ensure that samples from a given site were not contaminated with remnant grains from a different location or experimental run.

The grain size distribution of each bed sample was measured using a Micromeritics SediGraph 5120. The SediGraph is a reliable instrument for measuring the grain size distribution of fine-grained sediments (e.g., Bianchi et al., 1999) and has been used extensively for SS proxy applications. The SediGraph calculates grain size distribution using x-rays to measure sediment concentration and settling velocity, which is then used to compute grain size using Stokes' Law:

$$d = \left(\frac{18 w_s \nu}{g'}\right)^{1/2} \tag{12}$$

Once bed samples were collected, water was decanted and bulk sediment was dried for 48 hr. Once samples were sufficiently dried, a 1.0 g split was separated for analysis in the SediGraph. The split was dispersed in 65 mL of analysis fluid: ultra-pure (18.2 Ω m resistivity) water for pure silt samples and 0.5% tetrasodium pyrophosphate (TSPP) ultra-pure water for samples containing clay, to reduce particle flocculation during analysis. Samples dispersed in TSPP underwent a 15 min ultrasonic bath to further eliminate flocculation of the sample.

### *2.6. Calculation of the mean sortable silt*

$SS_{mean}$ was calculated from the natural log transformed bin sizes, $\theta_i = \ln(d_i)$ where $d_i$ is the percent finer than bin edge in microns used in the process of measuring and quantifying the total grain size distribution of a sample (e.g. a SediGraph output). $SS_{mean}$ is defined as:

$$SS_{mean} = e^{\theta_m} \tag{13}$$

with $\theta_m$ defined as,

$$\theta_m = \sum_{i=1}^{n} \overline{\theta_i} f_{ss,i} \tag{14}$$

where $\overline{\theta_i}$ is equivalent to the "natural log of the bin's geometric mid-point diameters", $\overline{\theta_i} = (\theta_i + \theta_{i+1})/2$ (McCave and Andrews, 2019), and $f_{ss,i}$ is the fraction of material associated with the SS size range (10 to 63 $\mu$m) in bin $i$. $i = 1$ is associated with the bin whose lower edge is 10 $\mu$m; $i = n$ is associated with the bin whose upper edge is 63 $\mu$m: $f_{ss,i} = p_i / \sum_{i=1}^{i=n} p_i$.

### *2.7. Defining the experimental conditions*

A series of preliminary experiments were run to determine the general repeatability of the methods and the influence of inlet sediment concentration, flow depth, and experimental runtime on the $SS_{mean}$ in the deposit for the purpose of defining the final experimental conditions. Repeat 2 hr experiments, even at different concentrations, generally showed less than $\pm 1$ $\mu$m variation between any individual $SS_{mean}$ statistic at a given sampling location. This suggests that the experimental methods show little variation in $SS_{mean}$ from run to run. The one exception to this was that up to 1.4 $\mu$m variation was observed at the most distal location (Fig. 5A). The fact that the variability between $SS_{mean}$ at a given location under different inlet concentrations (C $\approx$ 500 and 1000 mg/L) was also in the same variation range indicates that the deposit grain size distribution is not dependent on the inlet concentration. Inlet concentration does significantly influence the deposition rate, but it does not fundamentally change the size distribution in the deposit. This result is advantageous because it allows for running experiments at concentrations higher than would be expected in a natural boundary current to speed up time without altering the size distribution in the deposit. It also means that variations in concentration at the inlet should not impact the results. The lack of dependence of grain size in the deposit on source concentration is also helpful for modelling since initial concentration data following an episodic resuspension event is sparse.

This study also examined the role of flow depth, $h$, on the deposit grain size distribution and resulting SS statistics (Fig. 5B). Unlike concentration, changes in flow depth do produce significant differences in $SS_{mean}$ at a given distance downstream of the input all else being constant (e.g., constant velocity and input sediment size characteristics). This is to be expected given the dependence of $C_i = C_i(x)$ on the ratio $\alpha w_{s,i}/(hU)$ (Eq. 9). In the simple depositional model, $h$ has as significant of influence on the concentration profile as $U$. In general, increasing flow depth coarsens the deposit at a given location due to the increase in the advection length scale of a given particle size $L_i = hU/(\alpha_i w_{s,i})$ (Fig. 5B). Because this study is interested in the relationship of $SS_{mean} = SS_{mean}(x, U)$, depth was held constant at $h = 3.4$ cm in all other experiments. Using this depth, the experimental Reynolds number, $Re = Uh/\nu$ ranged from 1500 to 9000 (low but turbulent).

If particles that deposit from the current to the bed do not move, then the grain size distributions in the beds will be independent of experimental duration. However, in these experiments, some of the deposited particles did move as bed load, and the downstream movement intensified with increasing velocity. The presence of bed load suggests that the size of the sediment in the beds at a given distance down channel could change with the total runtime of the experiment. The sensitivity of the measured $SS_{mean}$ to variations in experimental run time was tested by comparing results between a standard experimental run time of 2 hr to an experiment with a run time of 6 hr (Fig. 5C). The differences in $SS_{mean}$ between these two experiments at the three upstream sampling locations all fell within the range of experimental variability. However, slight coarsening of $SS_{mean}$ ($\approx 1$ $\mu$m) at the two most distal locations was observed (Fig. 5C). This is attributed to downstream transport via bed load motion of larger grains sizes.

Given that experimental run time could impact the distribution of SS, the run time for all experiments was fixed at 2 hr and varied the inlet concentration to ensure that enough sediment deposited during the 2 hr to be sampled and sized with the SediGraph. Flow depth was fixed at 3.4 cm for all runs, and velocity was varied from 5 to 25 cm/s (Table 1). Taken together, these conditions and procedures enable isolation of the link between $SS_{mean}$, flow velocity, and distance from the input for each of the four sediment mixtures. The scaled distance over which the deposits in the flume develop is from $x/h = 0$ to $\approx 260$, which is equal to a corresponding distance of 0 to 26 km for a 100 m thick boundary current.

## 3. Results

### *3.1. Transport conditions*

The key hydraulic and sediment transport parameters associated with each of the five experimental velocities are first presented for the purpose of contextualizing the conditions before describing the bed morphology or $SS_{mean} = SS_{mean}(x, U)$ results. The flow parameters presented are: the shear velocity, the bed shear stress, a measure of the thickness of the initial viscous sublayer, $\delta = 5\nu/u_*$, and the Saffman parameter, $S$. Two other sediment transport ratios are also provided for silt sizes 10, 30, and 60 $\mu$m. The two ratios are $u_*/w_s$, a measure of how well mixed particles in suspension are over the vertical, and $\tau^*/\tau^*_{cr}$, a measure of particle mobility or transport intensity for those particles that make it to the bed. For the transport intensity parameter, $\tau^*$ is the dimensionless bed shear stress ($\tau^* = u_*^2/u_g^2$) and $\tau^*_{cr}$ is the value of the dimensionless shear stress where significant bed load motion occurs. Here the $\tau^*_{cr}$ threshold was calculated as $\tau^*_{cr} = [0.22Ga^{-0.6} + 0.06\exp(-17.77Ga^{-0.6})]/2$ where $Ga$ is the Galileo number, $Ga = u_g d/\nu$ (a type of particle Reynolds number, (e.g. Charru and Hinch, 2006)), with $u_g$ being the particle velocity scale associated with the submerged gravitational body force, $u_g = \sqrt{g'd}$ (García, 2008). The equation for $\tau^*_{cr}$ yields the classic Shields curve divided by 2. This reduced Shields threshold was chosen since it predicted motion for cases where

$U \geq 15$ cm/s, which is in line with the observations in this study. All of the contextual parameter and ratio values are given in Table 2.

Values of $u_*/w_s$ indicate transport in the downstream direction being dominated by suspended load, $u_*/w_s > 1$, and, at least for the $d = 10$ and 30 $\mu$m cases, being fairly well mixed over the vertical, $u_*/w_s > 6$; which is equivalent to a Rouse number of $Z_R = w_s/(\kappa u_*) < 0.4$ (García, 2008). Nevertheless, all runs except for the case of $U = 25$ cm/s with pure clay experienced deposition and the development of a bed that could be sampled at at least two locations in the flume.

In all runs, the thickness of the viscous sublayer exceeded the diameter of the silt in the mixture by a factor of roughly 10 (i.e., $\delta > 400$ $\mu$m). This suggests that any sediment that made it to the boundary would be submerged in the region of flow dominated by viscous effects. In such cases, the likelihood of re-entrainment of particles from the wall region can be quantified with $S$ (Saffman, 1965; Hamm et al., 2009); Saffman (1965) proposed a value of $S = 0.65$ is needed for the average viscous lift force to overcome the gravitational force on a particle. The highest value of $S$ was $\approx 0.15$. The authors therefore expect that most of the particles that make it to the bed will likely remain in the near bed region rather than being resuspended up into the flow.

While the Saffman values suggest that experimental conditions are in line with a net depositional setting for silt, it is still possible for particles on the bed to move in the downstream direction as bed load. The ratio of $\tau^*/\tau^*_{cr}$ provides a measure of the flows ability to move particles on the bed with values greater than 1 suggesting motion. In all experiments with silt, the onset of bed load was found to be well captured by this ratio. For runs with $U < 15$ cm/s, the bed that developed was largely immobile and topographically featureless. However, for all runs containing silt in the input sediment and $U \geq 15$ cm/s, active bed load occurred throughout the experiment and led to the development of migrating bedforms similar to those of Mantz (1978), Hollister and McCave (1984), Hamm and Dade (2013), and Yawar and Schieber (2017).

### 3.2. *Bed morphology*

The deposit thickness, morphology, and grain size all varied spatially down the flume and with flow velocity. A constant spatial trend in all cases was that the deposit thickness decreased in the downstream direction. When silt was present and $U \geq 15$ cm/s, the bed morphology transitioned downstream from a flat deposit with a few moving particles in the surface layer, to migrating two-dimensional (2D) ripples, and then to migrating barchan shaped three-dimensional (3D) ripples. A similar patter of a featureless deposit, to 2D ripples, to barchan ripples was also observed at a given station with increases in velocity. Both trends in morphology are interpreted as an outcome of a reduction in the amount of silt in the bed load layer with distance from the inlet and with increased velocity at a station. Input material also influenced the type and size of bedforms present. Increases in clay content moved the transition from a flat bed to 2D ripples, or from 2D ripples to barchan ripples, farther downstream (Figures 6-10).

At a flow velocity of 5 cm/s (Fig. 6), no bedforms were observed with any of the four inlet sediment mixtures. Instead, deposition resulted in a smooth, uniform bed throughout the length of the flume. At a flow velocity of 10 cm/s (Fig. 7), bedforms are minimal with some small ($< 1$ cm wavelength) 2D ripples being apparent in the pure silt runs, but no true ripples forming in any run with clay.

At $U = 15$ cm/s (Fig. 8), 1-2 mm tall ripples were present at the first three sampling locations in all experiments containing silt; ripples were not present in the pure clay experiment. The pure silt experiment saw a continuation of these ripples to the end of the flume, while the experiment with clay saw the ripples transition into barchan ripples around 6.4 m for the 1:1 experiment, and around 5.5 m for the 1:2 experiment.

At $U = 20$ cm/s, pure silt beds transitioned from 2D ripples to barchans moving downstream (Fig. 9). A similar pattern occurred for the clay and silt experiments, with the structures becoming more disorganized as clay content increased. Pure clay resulted in no bedforms with only a few bare patches in the bed followed downstream by a zone of no net accumulation of sediment. Bed load did occur in the form of what have been called floccule streamers (Schieber et al., 2007) in the zone of no net accumulation (Fig. 9, lower right). The streamers convey sediment parallel to the flow direction within low-speed streaks and were visually evident in all experiments. Increasing velocity to 25 cm/s (Fig. 10) resulted in an increase in height of the pure silt barchans (up to 8 mm, $h/h_{bedform} = 4.25$). Bedforms in the clay and silt mixtures became increasingly sparse. Large clay aggregates, likely an artifact of the mixing process, were also mobile at this velocity and were able to move as bed load down the flume, some of which deposited in the stoss and lee sided of the ripples. A thin deposit of these clumps was also observed from 0-3.8 m along the flume. The experiment with pure clay did contain bed load transport but yielded no net deposition.

### *3.3. Downstream patterns in deposit grain size*

All of the experiments using silt, regardless of the amount or type of clay added, exhibited some degree of systematic downstream fining of the SS fraction in the bed (Fig. 11). Downstream fining was stronger (more change in the bed size distribution for each step in distance downstream from the inlet) for the lower velocities than it was for higher velocity (Fig. 11). In fact, for some of the 25 cm/s runs, little change was observed in the deposit grain size distribution. This result could be interpreted as the system either needing longer distances to observe the fining or the system pushing more towards an equilibrium state between deposition and re-entrainment as velocities increase.

The size distribution data in Figure 11 also shows that the distribution of SS at a given station coarsens with current velocity. This can be observed by comparing the cumulative distribution curves at $x = 4.57$ m in the top row of the figure for velocities of 5, 15, and 25 cm/s. Because the conditions are overall net depositional ($E_s \approx 0$), the coarsening with the increase in velocity can be interpreted as the outcome of an increase in the advective length scale, $L_i$, (Eq. 10).

### *3.4. The relationship between sortable silt and current velocity at a station*

$SS_{mean}$ was calculated following the method described in the *Calculation of the mean sortable silt* section for every bed sample for which silt was present in the input sediment (i.e., for all runs except the pure clay runs). The range of $SS_{mean}$ across all experiments and station locations was 15 to 45 $\mu$m. $SS_{mean}$ systematically decreases progressing from the flume entrance to exit following a trend similar to the $d_{50}$ of the SS fraction (Fig. 11).

All $SS_{mean}$ data were grouped by station and plotted against current velocity to examine the relationship between $SS_{mean}$ and $U$ (Fig. 12). Linear regression was performed with $SS_{mean}$ as the scalar response and $U$ as the explanatory variable at each station. The regression was performed for data for each sediment type independently and also for a combined dataset using $SS_{mean}$ values from all three sediment mixtures. The fit equations for the combined dataset are shown in Figure 12; the coefficients and $R^2$ values for all regressions are given in Table 3. Of the four sets of regression coefficients, those obtained from the combined dataset have been chosen here (i.e., Fig. 12 and the "All" rows in Table 3) to be the most significant since they were developed with the largest number of data points. For these regressions, $R^2$ ranges from 0.7 to 0.94 (Table 3).

Three trends are evident in the regression output. The first is that the data tend to be better described by linear regression as the distance from the flume inlet increases (Fig. 12). The second and third are that the slope of the fit line increases with distance from the inlet (from 0.54 to 0.8) and the intercept decreases (from 23.8 to 13.2).

While not shown in the figures or tables, a regression was also performed between $SS_{mean}$ and the shear velocity, $u_*$. For this particular set of data, no predictive power was gained by using $u_*$ instead of $U$. For this reason, and because the classic SS hypothesis relates $SS_{mean}$ and velocity, the discussion focuses on the relationship between $SS_{mean}$ and velocity rather than shear velocity or bed shear stress.

## 4. Discussion

Before discussing the results, it is important to highlight that this study only has the ability to examine $SS_{mean} = SS_{mean}(x, U)$ under the action of downstream advective sorting in a depositional environment with minimal to no bed load transport. Furthermore, because the flocculation potential of the sediment mixture was low, the results presented reflect conditions unaffected, or minimally affected, by flocculation. Additional discussion of the study limitations and results context are given below following the discussion of the results.

### 4.1. Comparison with field and other laboratory data

A primary outcome of the study is the experimental demonstration that silt in a depositional system does sort advectivally both downstream and at a station as a function of velocity over the range of $U = 5$ to 25 cm/s. Within this context, it is found that grain size sorting is dependent on the distance from the flume inlet to the sample location and the flow depth; an additional analysis discussed below also shows that it is dependent on the size distribution of the sediment at the source or inlet location. Furthermore, under the present experimental conditions, $SS_{mean}$ appears to be linearly, or nearly linearly, related to the average current velocity at a particular distance from the flume inlet.

The studies of Ledbetter (1986) and McCave et al. (2017) provide the only known field data examining the relationship between the mean silt (Ledbetter, 1986) and SS (McCave et al., 2017) and current velocity. Even with the current velocity measured at a variety of locations and the natural complexity inherent in a field site, both of these studies found that a measure of the average silt size in the deposit was related to the current velocity. For example, the data of Ledbetter (1986) yields a relationship of $d_{ms} = 0.46U + 13.95$ ($R^2 = 0.82$) where $d_{ms}$ is the mean silt size in $\mu$m and $U$ is the current velocity in cm/s. Here the slope, 0.46, is only slightly lower than the values obtained with the laboratory study herein with the intercept falling at the low end of the measured range (Fig. 12). Furthermore, in McCave et al. (2017), slope values from the regression of $SS_{mean} = mU + b$, with $SS_{mean}$ in $\mu$m and $U$ in cm/s, ranged from $m = 0.59$ to 0.88 ($\mu$m-s/cm) with intercept values between $b = 15.6$ to 7.6 ($\mu$m). These slope values from McCave et al. (2017) are all inline with those obtained in the present flume study, with the intercept values being slightly smaller than those herein.

The similarity in form of the relationship and the values of the regression coefficients between this laboratory study and the field suggest two co-supporting lines of thought. The first is that the laboratory experiments reasonably reproduced advective depositional sorting even at their reduced scale. The second, assuming the flume experiments do capture the first-order physics, is that the sorting experienced in the field might be an outcome of downstream advective sorting. If this is the case, then it would also be reasonable to suggest that changes in the slope or intercept values along a deposit could be reflective of the distanced from a suspension source location.

To the best of the authors' knowledge, there have been no laboratory studies that have examined the relationship between $SS_{mean}$, $U$, and distance from the suspended sediment source. The closest studies to this one have been the studies of Hamm et al. (2009) and Hamm and Dade (2013). Both of these studies examined the dynamics of silt transport in a recirculating racetrack flume. However, only Hamm and Dade (2013) specifically set out to examine how the grain size distribution and $SS_{mean}$ varied with current velocities ranging from 20 to 53 cm/s. For the experiments, those authors used glass microspheres with diameters in the range of 13 to 44 $\mu$m. Similar to the observations here, the experiments produced both longitudinal streaks of clustered silt particles moving as bed load and mobile barchan shaped ripples. However, contrary to the present findings, the study did not report evidence of grain size sorting within the bedforms as a function of flow velocity.

The authors suggest that there are at least three differences between this study and that of Hamm and Dade (2013) that likely account for the different outcomes of the two studies with respect to sorting of silt. The first is that the velocities used herein are nearly all lower than those of Hamm and Dade (2013). In the current experiments, velocity ranged from 5 to 25 cm/s compared to the 20 to 53 cm/s used by Hamm and Dade (2013). The second major difference is that this work used a flow through flume rather than a racetrack flume, which enabled examination of the role of downstream advective sorting. The third is that crushed silica silt and clay mineral was used as the sediment type instead of glass microspheres. The authors expect that the first two differences are the most significant in driving the differences in bed texture relationships with current velocity between the two studies.

### *4.2. Modeling the trends in* $SS_{mean} = SS_{mean}(x, U)$

This section explores the ability of the simplified model to capture the trends in the experimental data, both in terms of downstream and at-a-station trends, and the model is used to consider how changes in the input sediment size distribution can impact

$SS_{mean} = SS_{mean}(x, U)$.

### 4.2.1. *Model comparison with downstream flume data*

If the equations for $\alpha_i$ and $E_{s,i}$ suggested by Hamm et al. (2009) (Eqs 3 and 4), and the Ferguson and Church (2004) settling velocity relation are used, then the simplified depositional model coupled with measured data has no calibration parameters. The inputs for the model include size distribution and inlet concentration (used to define $C_{i,0}$), the flow depth, velocity, and shear velocity, with the resulting output being $f_{b,i} = f_{b,i}(x, U)$ from which $SS_{mean}$ can be calculated. Of the measured inputs, those related to the hydraulics are well constrained with little experimental or measurement variability. Measured silt distributions from a deposit location under identical hydraulic conditions varied little from run to run in the preliminary experiments. Nevertheless, some variability was observed in the shape of measured size distribution of the silt in the sediment hopper from run to run. Because the depositional model is sensitive to changes in the initial grain size distribution over the depositional length of the flume, the authors elected to compare the model to data from the flume using a synthetic initial SS size distribution. The synthetic distribution was based on an average of the measured values and assumes a normal distribution of log transformed grain sizes (as described in Section 2.2).

The model captures the general shape of $SS_{mean}$ with distance from the source for a given velocity. The model does the best at capturing the data for the intermediate velocities ($U$ = 10-20 cm/s). For low velocities ($U$ = 5 cm/s) the model generally predicts a coarser $SS_{mean}$ than the observation, whereas for the higher velocities it tends to underestimate $SS_{mean}$ relative to the experimental data (Fig. 13).

One reason for the underestimation of $SS_{mean}$ at higher velocity might be the size independent nature of $\alpha$ and $E_s$ in the Hamm et al. (2009) formulations and the near zero value of $E_s$ for the particular flow conditions here. Rather than being size independent, one could expect $E_s$ to go up as grain size reduces under the consideration that it is easier to entrain smaller particles. This reasoning is common in entrainment functions for sand (e.g., Garcia and Parker, 1991), but it goes against the reasoning and data presented in Hamm et al. (2009) for entrainment of particles smaller than the viscous sublayer thickness. If the erosion and deposition functions did include an element of size dependence, finer material would stay in suspension longer at higher flows relative to large particles.

An attempt to improve the model to better match the experimental data was done by introducing the approach of Mooneyham and Strom (2018) to model the erosion and deposition terms (in place of Eqs 3 and 4). The Mooneyham and Strom (2018) approach is less sophisticated than other deposition and entrainment functions, but it does provide size-class-dependent net deposition and erosion functionality and it was developed for suspensions of clay and silt moving over impermeable and permeable beds. However, even after tuning the coefficients in the model, the method was not able to provide a substantial increase in descriptive power over the model described in the *Depositional theory* section using Eqs 3 and 4 for $\alpha$ and $E_s$.

It was found that the best match between the flume deposit data and the model was obtained by altering the size distribution of the inlet sediment. Reasonable matches between the data and the model could be obtained by increasing the mean size of the inlet distribution with current velocity. For example, for the pure silt case, using $\theta_m =$ 2.6, 2.8, 2.8, 3.2, and 3.5 for the cases of $U = 5$, 10, 15, 20, and 25 cm/s (with $\sigma = 0.8$) yielded a good fit between the model and data. An even better fit could be obtained by also changing $\sigma$ by up to $\pm 0.2$. Because the authors have no reason to expect that the inlet size distribution in these experiments was a function of current velocity, a single size distribution in the modelling analysis has been selected for simplicity.

*4.2.2. Trends in $SS_{mean}$ with velocity at a particular distance from the source*

Similar to the experimental data, this model predicts a general steepening of the $SS_{mean} = SS_{mean}(U)$ relationship (stronger sorting) with an increase in the distance from the flume inlet (Fig. 1B and 12); past the length of the flume the slopes lessens as coarser material is no longer available in suspension (Fig. 1B). While the trends in the experimental and model slopes move in the same direction, the model predicts a logarithmic form of the relation at the sampling location in the flume ($x/h = 43$ to 217 in Fig. 2) rather than a linear relationship. Near linear relationships are predicted farther downstream once the coarser material has fully settled out, but over the length of the flume the model is logarithmic.

The discrepancy in the shape of the relationship could be due to at least one of the following two explanations. The first is that it is possible that $SS_{mean} = SS_{mean}(U)$ is truly logarithmic at a given station, but that our experimental data were not able to capture the underlying functionality. That is, given the experimental variability, the limited number of velocities tested, and the near linear shape of the underlying log relations, the data were insufficient to differentiate between a linear and log form. The second possible explanation is that the model is too simplistic and either needs to account for changes in the arrival rate of sediment to the bed or re-entrainment of particles that make it to the bed.

#### *4.2.3. The role of input grain size distribution*

Modelling the downstream trends in grain size revealed that the model is dependent on the input grain size distribution, which in natural systems can, in turn, be related to the ultimate source of the sediment. The potential for the relationship between $SS_{mean}$ and current velocity to be dependent on the size distribution of the source sediment has been discussed in the literature (McCave and Hall, 2006). Indeed, McCave et al. (2017) attributed differences in the linear regression coefficients between $SS_{mean}$ and $U$ they observed in different groupings of data to differences in the nature of the input sediment and potentially to differences in the distance from the source.

The model is used here to examine how changes in grain size distribution of the input material impact the $SS_{mean} = SS_{mean}(U)$ relationship at three different downstream distances (Fig. 14). The distances from the source are given in terms of location scaled with depth, $x/h$. The position $x/h = 217$ corresponds to the most downstream sampling location in the flume using the experimental depth. The two additional locations are roughly double that and ten times that. For a 100 m thick current, the three positions would correspond to 22, 50, and 200 km from the suspension source. The analysis confirms that the $SS_{mean} = SS_{mean}(U)$ relationship, in an advective depositional sorting environment, depends on both the distance from the input and the size distribution of the input sediment. Coarsening of the input, and moving up-current so as to be closer to the source, both have the effect of producing deposits that are relatively coarser. Coarsening of the input sediment also leads to an increase in the slope of the $SS_{mean} = SS_{mean}(U)$ relationship (stronger sorting) at a given distance from the input. Figure 14 also suggests that differences in the $SS_{mean} = SS_{mean}(U)$ relationship between difference source material grain size distributions decreases with distance. It should be noted that changes in the source sediment size distribution with time could also lead to variations $SS_{mean}$ with depth in a deposit without any change in the current velocity. Variation in source sediment grain-size distribution and its potential confounding effects on interpretation of $SS_{mean}$ is also discussed in McCave et al. (1995) and McCave and Hall (2006).

### *4.3. Study context and limitations*

#### *4.3.1. Critical conditions*

The presented experiments show that silt can deposit at velocities of at least 25 cm/s ($u_*$=1.38 cm/s) with a $d_{50}$ of around 35-40 microns. This value is significantly larger than the critical depositional shear velocity of 0.67 cm/s suggested in McCave et al. (2017), but is inline with the results of other laboratory experiments using silt such as Mantz (1978), Hamm and Dade (2013), and Yawar and Schieber (2017). Hamm and Dade (2013) saw silt deposit into barchan ripples at velocities up to 30 cm/s (0.28 Pa) and Yawar and Schieber (2017) observed barchan ripples form at velocities of up to 55 cm/s (material $D_{50}$ of 50 microns), and 40 cm/s (material $D_{50}$ of 30 microns). Similar barchan shaped bedforms have also been observed in deep sea deposits (e.g., Hollister and McCave, 1984). Based on these values and observations, it seems reasonable to expect that silt can actively deposit to the ocean floor under the majority of boundary current conditions.

Deposition of clay on its own seems to achieve a critical velocity between 20 and 25 cm/s (0.13 - 0.19 Pa), although the inclusion of silt appears to increase this point, potentially due to low energy areas around silt bedforms. McCave et al. (1995) cites a critical shear velocity measured in a radial laminar flow cell by Self et al. (1989) for particles with a diameter on 10 microns to be 0.32 cm/s. The critical shear velocity and shear stress for the deposition and accumulation of clay in the experiments herein was around $u_* = 1.12$ cm/s and $\tau_B = 0.13$ Pa, higher than other reported values (e.g., McCave and Swift, 1976; Self et al., 1989; McCave, 2008). Clay deposition thresholds from the present experiments are inline with those of Schieber et al. (2007) and Yawar and Schieber (2017) where it was found that floccule ripples can form at velocities ranging from 10-26 cm/s. These high critical conditions for silt and clay deposition, and the consolidating effect of deposited clay, could lead one to conclude that only under high-magnitude events (e.g., benthic storms; Gardner et al., 2017) does erosion occur. In between such episodic events, contour currents likely function as a depositional system and advectively sort material that was resuspended during short-lived, high-energy events. However, the authors note that direct, observation-based knowledge of such fundamental processes in the deep sea is relatively limited, highlighting the importance for more field-based research of these systems.

#### *4.3.2. The role of clay content in the input sediment*

The presence of clay in the inlet sediment had an influence on the morphology of the bed that developed with time. The higher the percentage of clay, the more suppressed the silty bedforms. It is possible that this suppression of bedforms might have been due to the cohesive nature of the clay (Schindler et al., 2015). It is also possible that bedforms were not as pronounced in beds that developed from the clay-silt mixture relative to those with pure silt because of the overall reduced concentration of silt in input sediment under equivalent inlet concentrations.

While inclusion of clay did change the bed morphology, added clay (or clay type) in the input sediment did not have a strong influence on the sorting properties of the silt. Both downstream and at-a-station trends in $SS_{mean} = SS_{mean}(x, U)$ were relatively insensitive to the amount of clay in the input sediment. This behaviour can be seen in Figure 12 and Table 3 where slope and intercept values in the regression of $SS_{mean} = mU + b$ show little variability with the silt to clay ratio. The authors expect this outcome is a reflection of the little influence the suspended clay has on advective depositional sorting in these experiments.

The presence of clay has the possibility of influencing sorting of silt through at least three mechanisms. The first would be through binding of clay and silt particles into flocs within the suspension, thereby altering the settling velocity of both fractions. While the effect on the silt settling velocity was small relative to the effect of the silt on the clay floc settling velocity, Tran and Strom (2017) did find evidence of the silt settling velocity being slightly reduced if it is traveling in flocs rather than unflocculated. If sorting in the deposit is primarily an outcome of advective depositional sorting, then the binding of silt size material in mud flocs within the suspension, thereby altering the settling velocity of the silt, would be the primary way in which the presence of clay could impact the sorting of silt.

This study looked for the occurrence of clay-silt floc binding in suspension and on the bed using the camera system explained in Rouhnia and Strom (2017) and Mooneyham and Strom (2018) and in the mixing tank experiments outlined in the methods. In all samples, no indication of binding of the two fractions in suspension was evident. The silt existed as independent grains and the clay in visible aggregates roughly the same size as the silt or below the resolution of the camera ($\approx 15\ \mu$m); no large, low-density flocs such as those in Tran and Strom (2017) were found. The lack of silt binding in mud flocs within the present experiments suggests that additional work should be done in the future to better constrain the control that flocculation may place in downstream advective sorting of silt.

The second way in which clay could have impacted the depositional sorting of silt is through alterations to the deposition and re-entrainment rates ($\alpha$ and $E_s$ in the model). If deposited clay provided a measure of cohesion, it might lead to increased retention of fine silt particles that might have otherwise been resuspended back up into the flow leading to less-well-sorted silt. While such a situation seems plausible, the experiments do not support this line of reasoning. There was additional variability in the measured deposit silt size distribution for the runs with clay, but the overall slope and intercept regression coefficients from the runs with clay were similar to those from the pure silt runs regardless of the amount of clay added. The similar slope of the regression, and the near zero value for $E_s$, suggest that cohesion between the clay and silt near the boundary did not strongly influence the resuspension rate.

A third way that clay could have impacted sorting of silt in the deposit, and one that is not captured by the model, is through clay and silt interactions within the bed load layer.

*4.3.3. Bed load*

Both the SS hypothesis and the depositional model assume that particles do not move along the bed once they are deposited; i.e., no bed load transport. However, bed load was apparent in all silt and silt-clay experiments at or above $U = 15$ cm/s. It is this bed load that resulted in the creation of ripples that would migrate downstream with or without sediment in suspension. The pure silt ripples that formed in this study's non-recirculating flume closely resembled in size and shape the ripples that formed in the racetrack flume experiments of Yawar and Schieber (2017) (at 25 cm/s) and Hamm and Dade (2013) (at 20 cm/s). The silt-clay ripples were also similar is size, but contained some clay clumps. Similar silty clay ripples have also been observed in the field (Hollister and McCave, 1984).

While the movement of clay particles along the bed was observed in the pure clay experiments, the formation of migrating barchan ripples composed purely of clay floccules was not seen, as was observed in Schieber et al. (2007) for similar current velocities. A potential reason for the lack of migrating clay bedforms in the present study might be the relatively short length of the flume. The advective length scale of clay is significant, and it maybe that more flume length would have been needed to deposit sufficient clay material at those velocities to allow for the formation of larger clay floccules and migrating barchan clay ripples.

Regardless, observations of bed load and bedforms in these experiments and others all indicate that sediment deposited to the bed does have the ability to transport downstream below a critical shear stress for re-entrainment. This is acknowledged in the SS hypothesis. However, it is suggested that bed load and selective erosion in mud is diminished due to cohesion in the bed. Leaving selective deposition as the key sorting mechanism (McCave, 2008). The implication of bed load transport of deposited grains is that particles that originally deposited at one location under advective sorting could move to another location over long enough periods of time before becoming buried and incorporated into the sedimentary record at a particular location. On its own, such movement could alter the at-a-station relationship between $SS_{mean}$ and $U$ such as those examined in this study. Furthermore, having bed load movement of both clay and silt size material could further complicate the relationships. In studying deposits generated from mixtures of silt and clay moving as bed load, Yawar and Schieber (2017) found that the smaller fraction of silt ($d$ = 1-30 $\mu$m, a portion of which is part of the SS fraction) separated from the broader silt fraction, bound with the clay floccules, and then deposited with the clay floccules.

Bed load transport in general, along with the potential for clay-silt interaction within the bed load layer, could complicate the simple advective depositional sorting relationships explored and modelled in this study. Bed load transport is not accounted for in the model, and it is possible that the short length of the flume and short run time of the experiments minimized effects due to bed load transport and potential interactions between the clay and silt in the bed load layer. It must be acknowledged that neither bed load motion over long distances nor any shift in deposit location for the silt fraction between 1 and 30 $\mu$m due to transport within clay floccules is accounted for. Furthermore, neither these experiments or model account for the winnowing effects of erosion. As such, caution should be used when seeking to extend the present findings to more complicated field settings.

## 5. Conclusions

This study used a laboratory experiment and simple suspended sediment transport theory to investigate the relationship between mean sortable silt in a deposit, $SS_{mean}$, and average current velocity at the time of deposition in an advectively sorted system. The velocity used in the study ranged from 5 to 25 cm/s, velocities typical of deep-sea environments. Under these conditions, the laboratory currents were largely depositional. The relationship between $SS_{mean}$ and velocity was examined both at a particular distance from the input of suspended sediment (at a station) and as a function of total distance from the sediment input (downstream). Sediment used in the experiments consisted of crushed silica in the silt size range and different mixtures of the silica silt with clay minerals. The combination of the experimental methods and materials led to advective depositional sorting where silt sizes in the deposit fine with distance from the input and coarsen with increasing velocity and current thickness.

From the experiments, it was found that $SS_{mean}$ was linearly related to $U$ at a particular location in the flume; this outcome is similar to the field calibration of McCave et al. (2017). Regression between $SS_{mean}$ in microns and $U$ in cm/s produced fits with $R^2$ values between 0.7 and 0.94 and coefficient values similar to those from the field, even though the scales of the two studies are very different. In general, the slope of the $SS_{mean} = SS_{mean}(U)$ regression increased, while the intercept decreased, with distance from the input in both experiments and theory over the scaled length of the flume.

A model for $SS_{mean}$ in the deposit was developed using simple theory. For the experimental conditions, the model was able to reasonably describe the size distribution of silt in the deposit as a function of the input grain size distribution, the distance from the input, velocity, and current thickness. Based on this, the model was extrapolated to distances of up to ten times the length of the flume (or out to 200 km for a 100 m thick current). Doing so enables an illustration of the impact of distance from the source and source grain size distribution on the $SS_{mean} = SS_{mean}(U)$ relationship for a broader range of conditions than those explored experimentally.

Both the experiments and theory demonstrated the importance of the thickness of the current, $h$, in the sorting process. More specifically, the theory shows that the amount of silt in size fraction, $i$, within a deposit is strongly dependent on the ratio $x/L_i$; where $x$ is the distance from the source and $L_i$ is the advective length scale $L_i = hU/(\alpha w_{s,i})$. This shows that that $L_i$ is linearly dependent on both $U$ and $h$. Doubling either will produce a doubling of $L_i$.

This study shows that silt can advectively sort under depositional conditions, and it highlights how the distance from the input, flow thickness, and changes in the input grain size can alter the relationship. The similarity in the linearity between $SS_{mean}$ and

$U$ and the regressed slope and intercept values between the laboratory and field suggest that the field data of McCave et al. (2017) might have also been the outcome of advective sorting.

While the study explores advective depositional sorting of suspensions containing silt and clay, neither the theory nor experiments account for complicating interactions in these two size fractions that can arise due to flocculation. The experiments and model also do not account for winnowing due to erosion or long-range bed load transport of silt size grains that, if it occurs, has the potential to also complicate the $SS_{mean} = SS_{mean}(x, U)$ relationship.

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| Sediment | Inlet Concentration, $C_0$ [mg/L] | Velocity, $U$ [cm/s] |
|---|---|---|
| Pure Silt | 374-1660 | 5, 10, 15, 20, 25 |
| Pure Clay | 482-2070 | 5, 10, 15, 20, 25 |
| Clay & Silt (1:1 ratio) | 328-1654 | 5, 10, 15, 20, 25 |
| Clay & Silt (2:1 ratio) | 393-1871 | 5, 10, 15, 20, 25 |

Table 1: Experimental Matrix

| $U$ [cm/s] | $u_*$ [m/s] | $\tau_B$ [Pa] | $\delta$ [$\mu$m] | $S$ [-] | $u_*/w_s$ | | | $\tau^*/\tau^*_{cr}$ | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | $d_{10\mu m}$ | $d_{30\mu m}$ | $d_{60\mu m}$ | $d_{10\mu m}$ | $d_{30\mu m}$ | $d_{60\mu m}$ |
| 5 | 0.003 | 0.01 | 1632 | 0.002 | 49.4 | 5.6 | 1.5 | 0.2 | 0.2 | 0.2 |
| 10 | 0.006 | 0.04 | 930 | 0.012 | 86.8 | 9.8 | 2.6 | 0.6 | 0.5 | 0.5 |
| 15 | 0.009 | 0.08 | 641 | 0.038 | 125.8 | 14.3 | 3.7 | 1.2 | 1.1 | 1.0 |
| 20 | 0.012 | 0.13 | 491 | 0.085 | 164.3 | 18.6 | 4.9 | 2.0 | 1.8 | 1.7 |
| 25 | 0.014 | 0.19 | 413 | 0.143 | 195.6 | 22.2 | 5.8 | 2.9 | 2.6 | 2.4 |

Table 2: Conditions at a given flow velocity. $\delta$ is the thickness of the viscous sublayer. Values of $u_*/w_s$ and $\tau^*/\tau^*_{cr}$ are given for representative grain sizes of 10, 30, and 60 $\mu$m.

| Sediment | Station, $L$ [m] | Slope [$\mu$m-s/cm] | Intercept [$\mu$m] | $R^2$ |
|---|---|---|---|---|
| Pure Silt | 1.52 | 0.69 | 24.0 | 0.96 |
| Pure Silt | 3.05 | 0.85 | 16.4 | 0.99 |
| Pure Silt | 4.57 | 0.76 | 16.1 | 0.89 |
| Pure Silt | 6.10 | 0.82 | 13.6 | 0.93 |
| Pure Silt | 7.62 | 0.88 | 11.8 | 0.99 |
| Silt & Clay 1:1 | 1.52 | 0.40 | 25.9 | 0.84 |
| Silt & Clay 1:1 | 3.05 | 0.73 | 16.0 | 0.90 |
| Silt & Clay 1:1 | 4.57 | 0.70 | 17.6 | 0.94 |
| Silt & Clay 1:1 | 6.10 | 0.89 | 14.3 | 0.99 |
| Silt & Clay 1:1 | 7.62 | 0.34 | 20.6 | 0.69 |
| Silt & Clay 1:2 | 1.52 | 0.44 | 22.8 | 0.88 |
| Silt & Clay 1:2 | 3.05 | 0.33 | 23.0 | 0.64 |
| Silt & Clay 1:2 | 4.57 | 0.70 | 16.2 | 0.98 |
| Silt & Clay 1:2 | 6.10 | 0.82 | 13.0 | 1.00 |
| Silt & Clay 1:2 | 7.62 | 0.74 | 13.2 | 1.00 |
| All | 1.52 | 0.54 | 23.8 | 0.70 |
| All | 3.05 | 0.66 | 18.1 | 0.81 |
| All | 4.57 | 0.73 | 16.5 | 0.91 |
| All | 6.10 | 0.82 | 13.9 | 0.94 |
| All | 7.62 | 0.80 | 13.2 | 0.94 |

Table 3: Linear regression coefficients for $SS_{mean} = SS_{mean}(U)$. Regression slope and intercept values are given for individual sediment times (e.g., Pure Silt) and for the combination of all data from different sediment input (All). $R^2$ values of 1 indicate conditions were only two points were available due to the lack of deposition in some silt and clay runs.

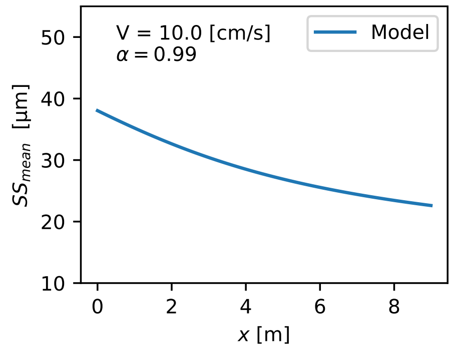


Figure 1: Solution of the simplified model for $SS_{mean} = SS_{mean}(x, U = 10 \text{ cm/s})$ based on Eqs 9 and 2

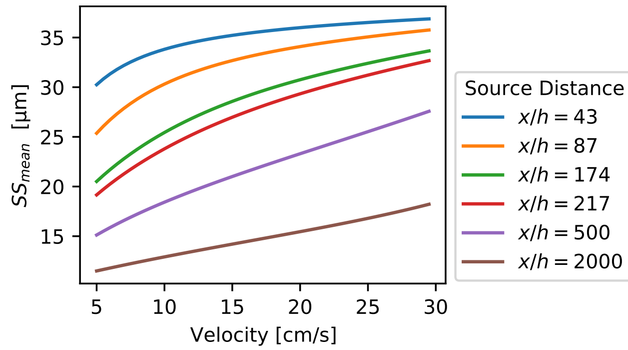


Figure 2: Solution of the simplified model for $SS_{mean} = SS_{mean}(x, U)$ based on Eqs 9 and 2. The each line shows the relationship between $SS_{mean}$ and $U$ at a particular distance from the source or flume inlet. The first four lines ($x/h = 43, 87, 174$, and 217) correspond to the sampling locations in the flume. The last two represent a hypothetical extension of the flume by a factor of 2 and 10 ($x/h = 500$ and 2000).

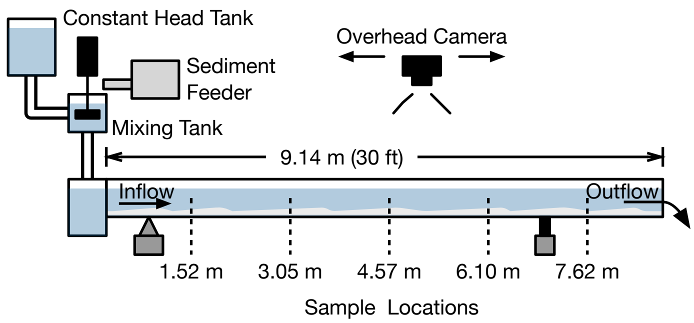


Figure 3: Experimental Setup.

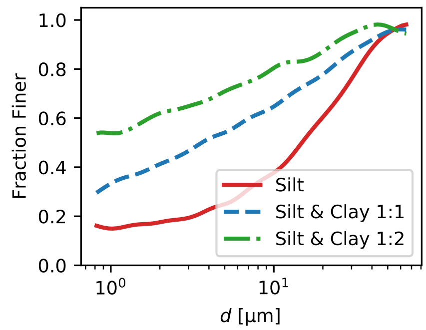


Figure 4: Initial grain size distributions for the three mixtures that included silt.

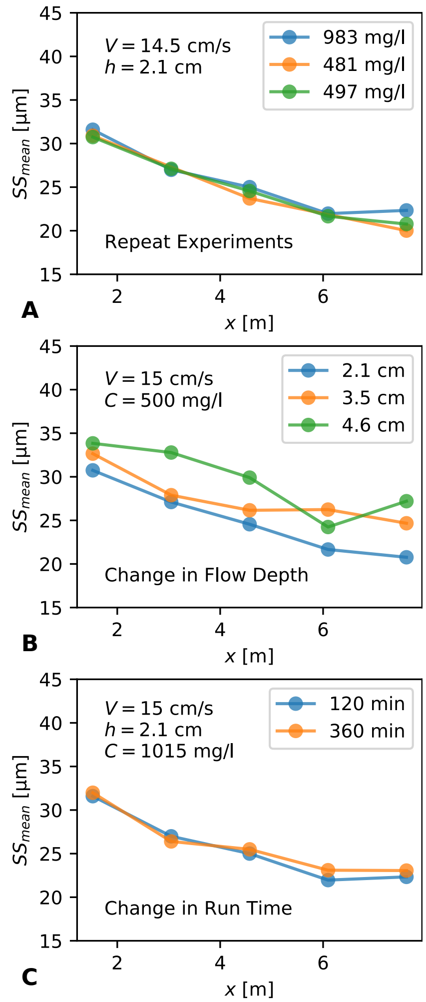


Figure 5: The distribution of $SS_{mean}$ down channel for the preliminary runs. (A) repeat experiments no change in experimental conditions except the inlet concentration, (B) changes in flow depth, $h$, only, and (C) changes in experimental runtime.

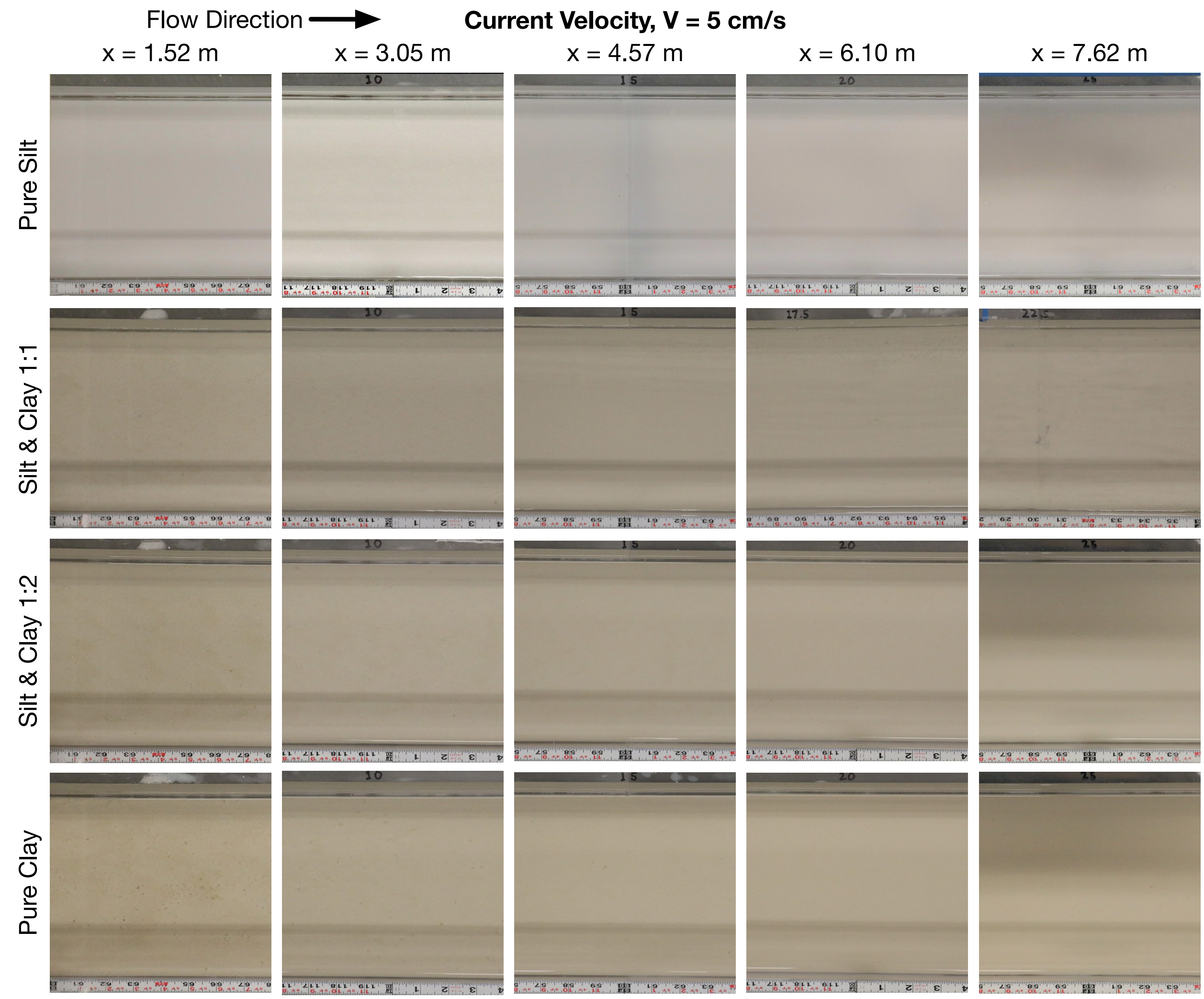


Figure 6: Images of the bed at the end of each of the $U = 5$ cm/s runs. $x$ is the distance downstream from the inlet.

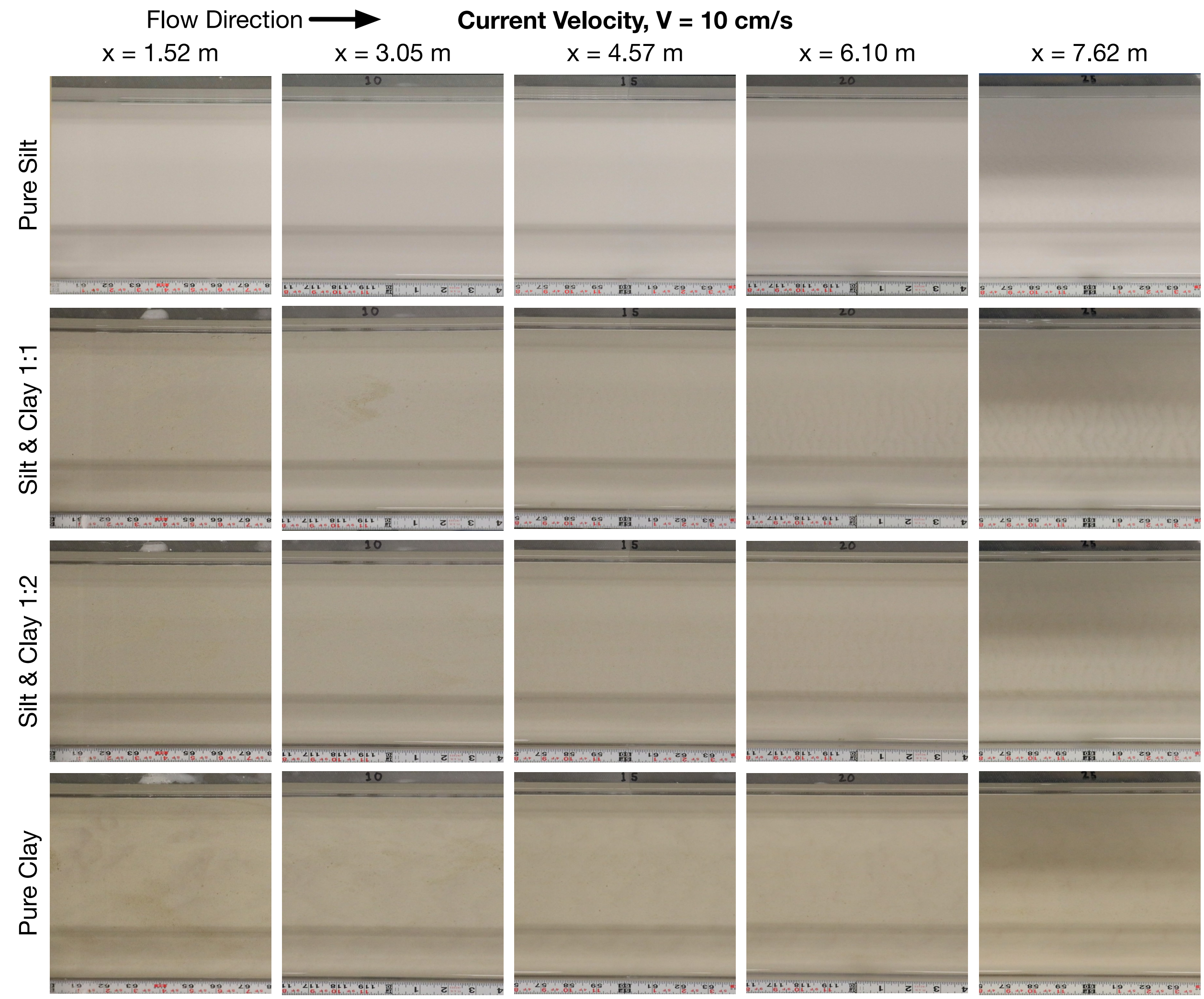


Figure 7: Images of the bed at the end of each of the $U = 10$ cm/s runs. $x$ is the distance downstream from the inlet.

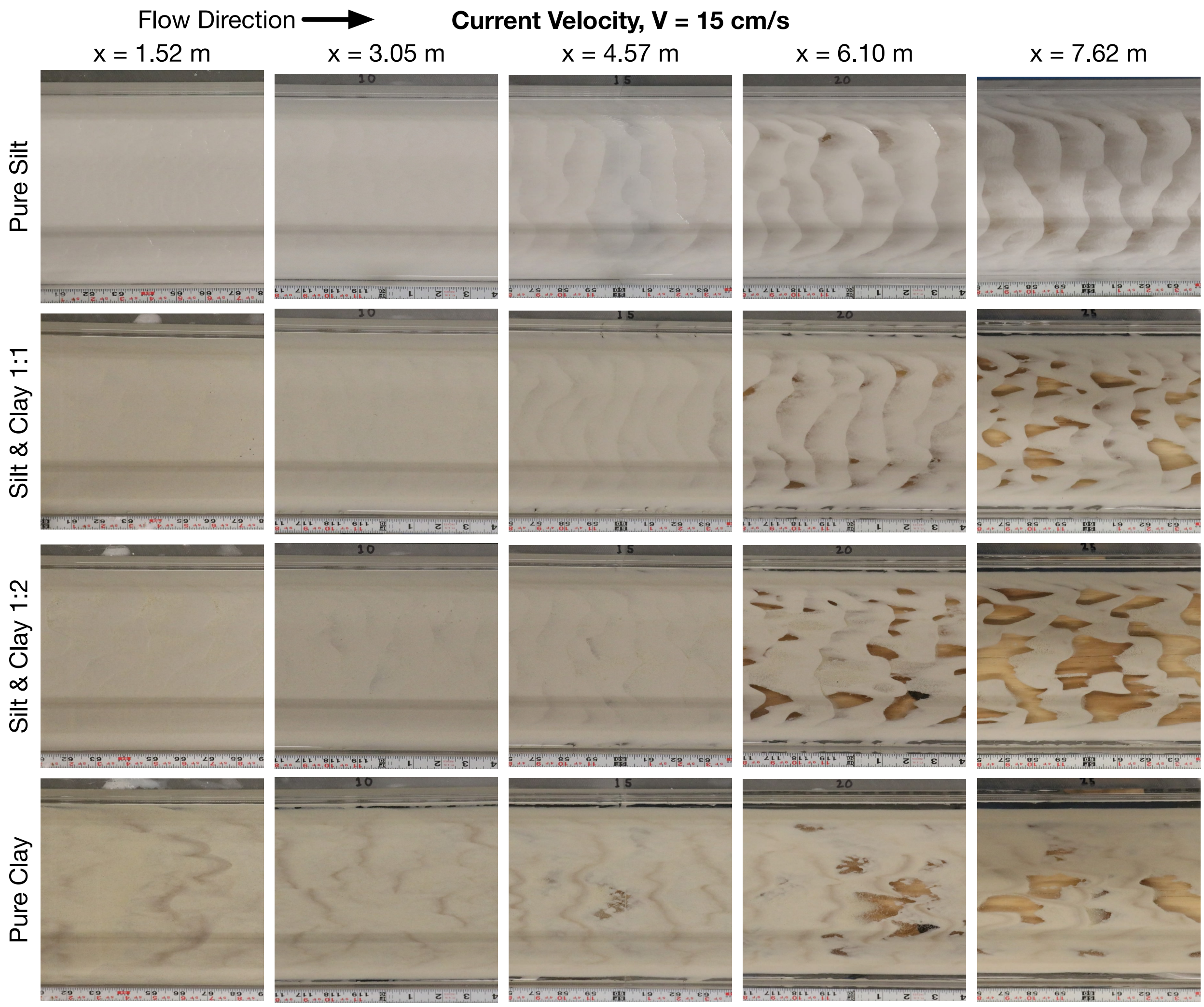


Figure 8: Images of the bed at the end of each of the $U = 15$ cm/s runs. $x$ is the distance downstream from the inlet.

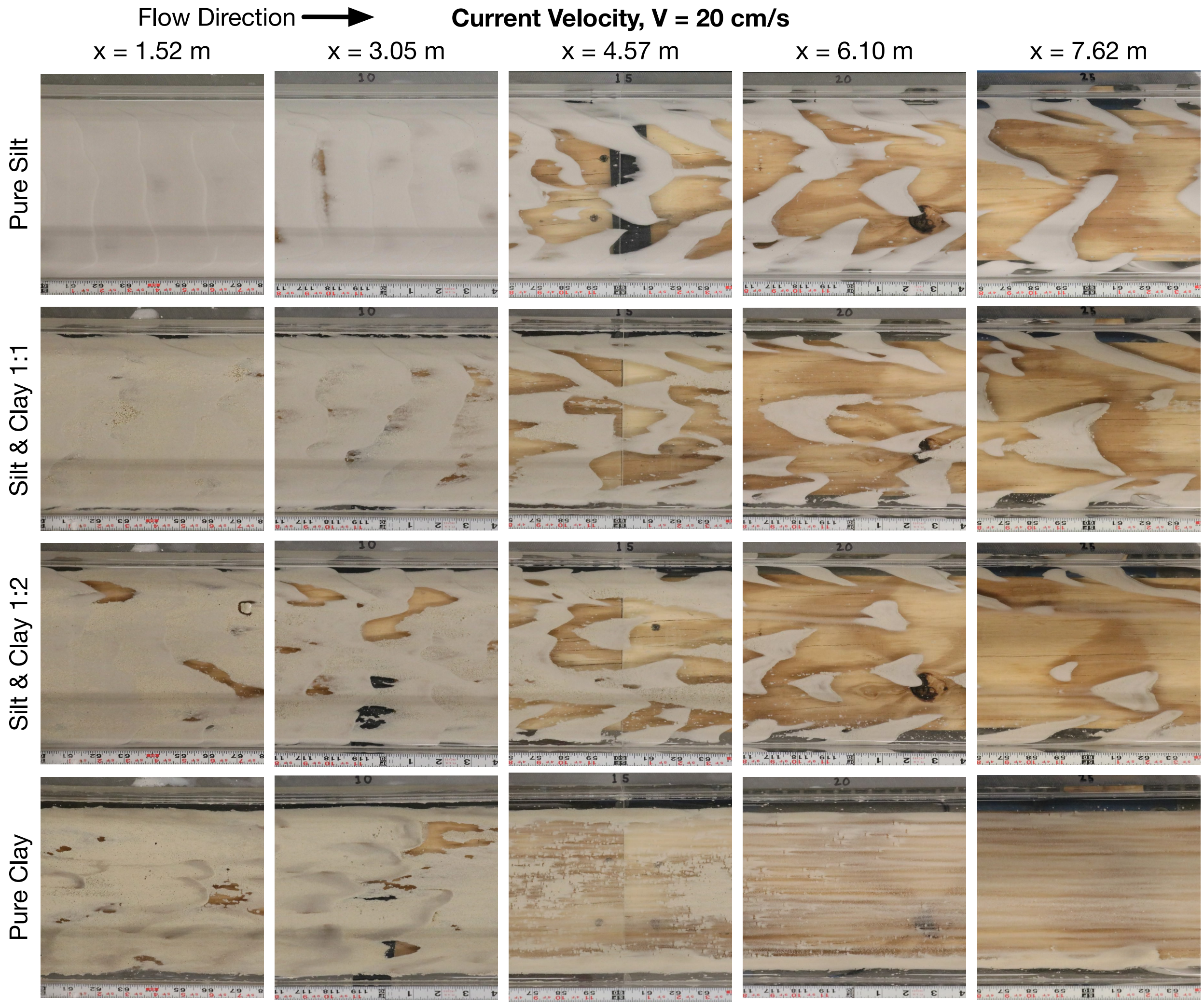


Figure 9: Images of the bed at the end of each of the $U = 20$ cm/s runs. $x$ is the distance downstream from the inlet.

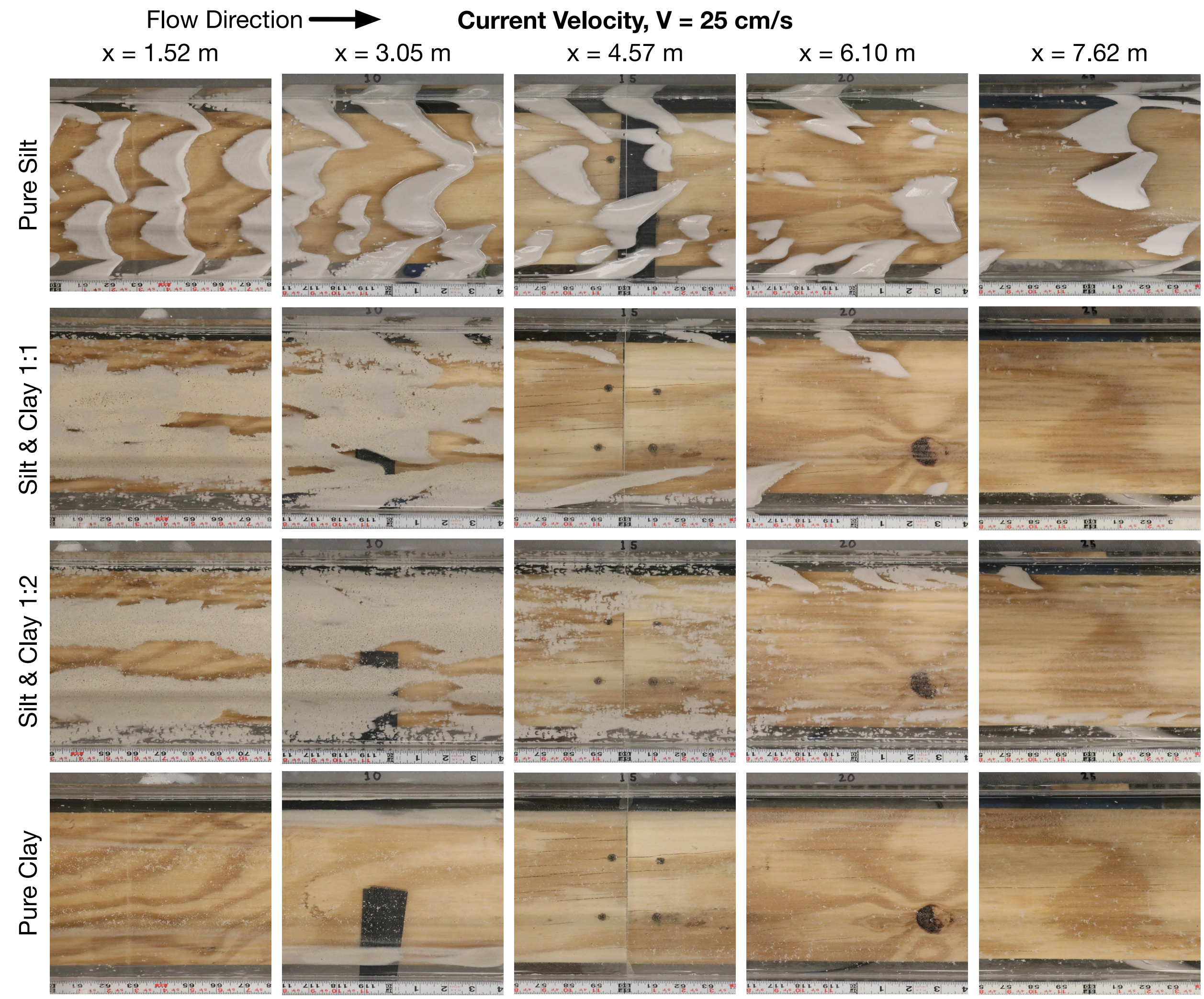


Figure 10: Images of the bed at the end of each of the $U = 25$ cm/s runs. $x$ is the distance downstream from the inlet.

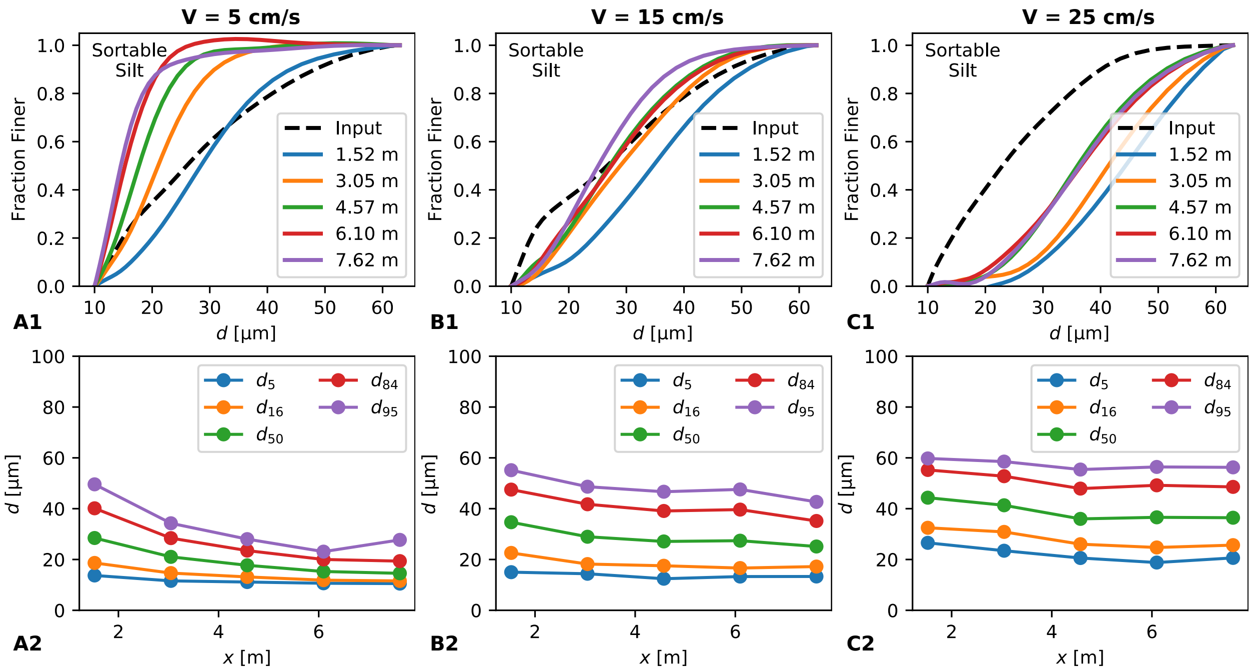


Figure 11: Downstream trends in the distribution of the sortable silt fraction for (A) $U = 5$ cm/s, (B) $U = 15$ cm/s, and (C) $U = 25$ cm/s; all plots are for runs with Pure Silt. At all velocities, the top row shows the cumulative distributions of the SS fraction and the bottom row shows the distribution statistics as a function of distance down channel from the inlet.

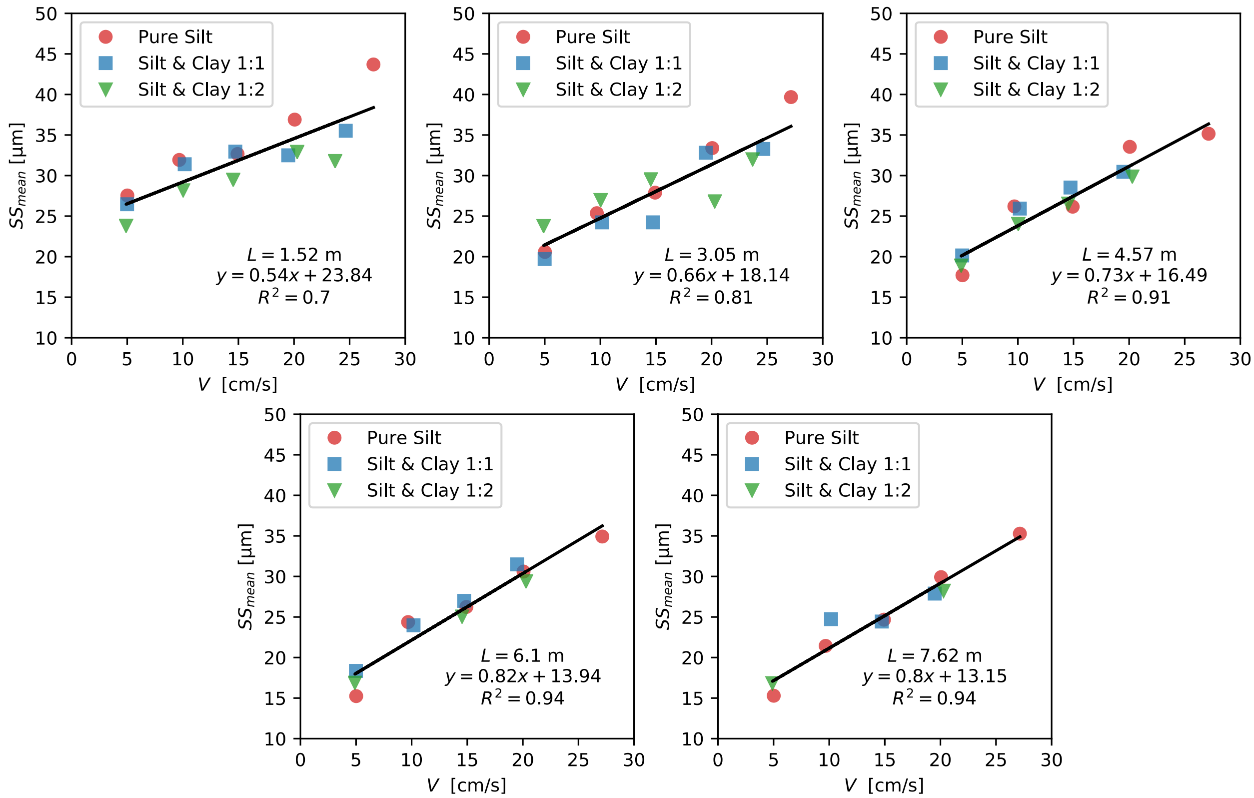


Figure 12: Regression between $SS_{mean}$ and current velocity at each of the five sampling locations. $L$ is the distance between the sediment input location and the sample location.

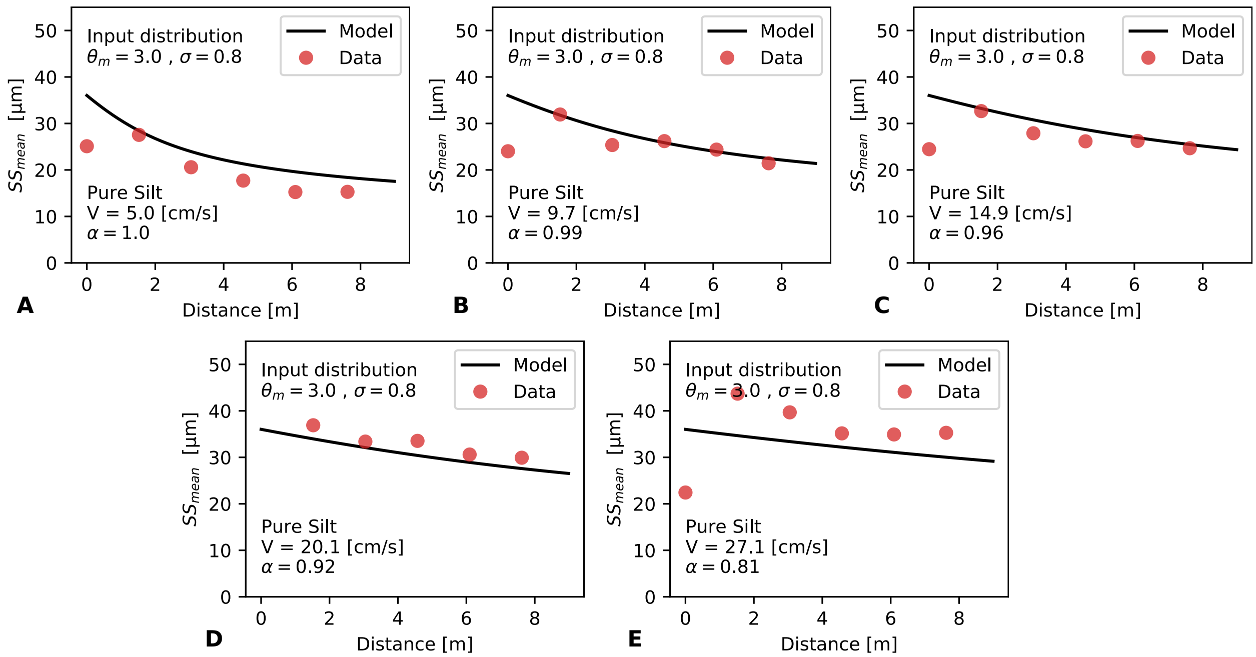


Figure 13: Comparison between the model and observations for low, moderate, and high velocities. $SS_{mean}$ at $x = 0$ m corresponds to the inlet sample $SS_{mean}$.

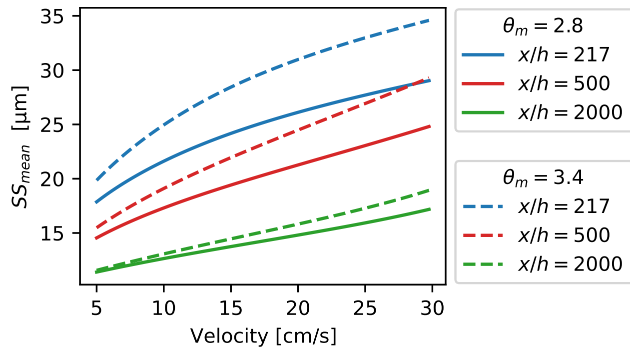


Figure 14: Model output showing the role of input grain size distribution on $SS_{mean}$ trends with distance from input and velocity. Colors represent distance from input. The two different line types represent the two different sediment input distributions. Input 1 (solid lines) is for a distribution with $\theta_m = 2.8$ (16 $\mu$m) and input 2 (dashed lines) has $\theta_m = 3.4$ (30 $\mu$m); both have $\sigma = 0.8$.