Abstract Prior studies demonstrate that bottom velocity measurements from Doppler sonar systems are proportional to bedload transport rates. These observations suggest that acoustically based systems offer a capability for rapid sampling of bedload transport processes. Before these measurements can be fully utilized, validation and understanding of the sampling mechanism are essential. We explore the measurement mechanism through a series of laboratory trials with a field instrument, the multi-frequency coherent Doppler profiler (MFDop). The MFDop system is a multi-frequency (1.2–2.2 MHz), bistatic Doppler sonar that provides three-component ensemble-averaged velocity profiles over a ~30 cm depth interval with up to 1 mm resolution at a rate of 50 profiles/sec. Tests of the MFDop system were carried out in the main flume in field-scale conditions at the St. Anthony Falls Laboratory (SAFL) using 1 m$^{-1}$ mean flows over a mobile bed of sand with median grain size $d_{50} = 0.4$ mm. We find agreement between MFDop transport measurements and measurements based on bedform migration rates, and sediment traps built into the SAFL flume. Predictions using the Meyer-Peter and Müller (1948) empirical equation closely match our observations while in contrast, predictions using the Nielsen (1992) equation are a factor of two higher.

1. Introduction

There is a long-standing and compelling need within the nearshore, continental shelf, and fluvial sediment dynamics communities for highly resolved space/time measurements of bottom stress and near-bed sediment flux. The need for improved and redundant estimates of bed shear stress—the main driver of sediment transport—is underscored by Grant and Madsen (1986) in their review of continental shelf bottom boundary layer dynamics. Nielsen (1992) summarizes sediment dynamics theory and knowledge of wave- and current-forced conditions, both separately and in combination; Van Rijn et al. (2013) and Lu et al. (2015) provide recent reviews of the topic. Models of coastal sediment transport are limited by the need for a better understanding of detailed transport processes over all sediment types and bedforms (Van der A et al., 2013) and improved observations of sheet flow and bedload transport (Van Rijn et al., 2013). And, the state of understanding of sediment transport in rivers is such that standard methods use models that are scaled by flow rates, and unsurprisingly, the resulting estimates are characterized by large variability (Gaeuman et al., 2015; Heyman et al., 2016). A key component to progressing in these fields continues to be the need for improved measurements of turbulence and sediment transport under field conditions. These observations are not easy to collect given the challenges of making observations in highly energetic environments without causing disturbance to the nearbed flow or erodible bottom sediments.

In the late 1980s and 1990s, acoustic sensing systems operating at MHz frequencies were proven to have the capacity for making high resolution measurements of suspended sediment concentration with minimum disturbance to the bed and the near-bed flow. These developments, which are reviewed by Thorne and Hanes (2002), include (a) the use of multiple acoustic frequencies to determine the vertical profiles of particle concentration and size under wave-forced conditions in the nearshore (Hay & Sheng, 1992) and in steady turbulent flow in a tidal estuary (P. D. Thorne & Hardcastle, 1997); and (b) the use of pulse-coherent Doppler sonar techniques for measuring turbulence-resolved profiles of velocity (Hurther & Lemmin, 1998) and of both velocity and particle flux (Shen & Lemmin, 1997, 1999; Zedel & Hay, 1999). And, it has been demonstrated that through careful use of commercially available Doppler sonar systems large-scale suspended sediment transport can be measured (see for example Conevski et al. [2020]; Guerrero et al. [2014, 2015]).
During the last two decades, the pulse-coherent Doppler approach has advanced to the point where it is now possible to obtain turbulence-resolving estimates of three-component velocities and particle concentration and size along a vertical profile at sub-cm resolution. These systems typically combine backscatter measurements used to determine sediment concentrations with bi-static beam geometries (sometimes called multi-static when two or more bi-static beam pairs are used) needed for velocity measurements, in some cases incorporating wide bandwidth transducers to enable the use of multiple frequencies (Dillon et al., 2011; Hurther & Lemmin, 2008; Hurther et al., 2011; Zedel & Hay, 2010). Importantly, the backscatter and velocity measurements are simultaneous in time and closely located in space, so that after inverting the backscatter to sediment concentration, estimates of the quasi-instantaneous particle fluxes are possible (The term “quasi-instantaneous” is used because some time averaging and spatial sampling are involved to reduce both the Doppler noise in the velocity estimates and the configuration/Rayleigh noise in the backscatter amplitudes.) Thus, while there remain important considerations related to system calibration and inversion methodologies—in particular for mixed sediment types and sizes—it is possible to state that, from a technological standpoint, the problem of measuring the flux of suspended, unimodal, sand-sized, non-cohesive sediment has been solved.

In contrast to flux measurements of suspended sediments, the challenge of measuring bedload transport in the field remains to be solved. The standard approach to such measurements remains mechanical sediment traps such as the Helley-Smith sampler that remain largely unchanged since the 1960s (D. W. Hubbell, 1964; D. Hubbell et al., 1985). These traps are highly invasive, can only provide time-integrated measurements of transport, and are hard to implement in energetic field conditions (Bunte et al., 2004; Marr et al., 2010). In cases where bedforms are present, they provide a means of estimating transport by repeated measurements to monitor their movements (Claude et al., 2012; Van den Berg, 1987). Simple tracers can provide an integrated record of overall transport and recently developed instrumented tracers can record details of individual tracer movements (see for example Pretzlav et al. [2020]). Approaches that attempt to measure transport instantaneously without disturbing the flow are possible based on acoustic methods. Passive acoustic (Geay et al., 2017), and seismic systems (Tsakiris et al., 2014) can indirectly detect the motion of sediment by recording the acoustic energy generated from particle collisions. More direct measurement is possible using active sonar systems. An early laboratory trial by Traykovski (1998) suggested that sonar techniques might be capable of indicating bedload transport. And, recent reports of Doppler profiling systems (ADCPs) used in sediment dynamics studies in rivers have demonstrated the ability of these systems to infer bedload velocity (but not transport) from the bias in the bottom-track velocity (Latosinski et al., 2017; Rennie et al., 2002; Ramooz & Rennie, 2010). Further, there have been successful comparisons of the apparent bed velocity to bedload transport rates estimated independently from bedform migration (Gauman & Jacobson, 2007) and recent laboratory trials by Conevski et al. (2019) demonstrate that after accounting for spatial sampling constraints, good correlation with bedload velocity can be achieved. Conevski et al. (2020) report on comparisons of bedload velocity measurements using three different ADCP types and four different operating frequencies; they find a correlation between measured velocities and bedload transport for all these systems but note significant differences depending on the frequency and system configuration. While these results are encouraging, the method is limited because (a) the acoustic beams in commercially available ADCPs insonify non-overlapping areas on the bottom, an issue when measuring over uneven bathymetry, as when bedforms are present; and (b) the thickness of the bedload layer is only partially resolved by these systems (depending on the frequency and acoustic configuration), and so bedload transport is not measured directly (see discussion in Conevski et al. [2020]). These results do however raise the potential for measuring the bedload component of transport building on the established methods for suspended load measurements. Indeed, such bedload measurements have provided encouraging results in recent laboratory studies utilizing purpose-built sonar systems (Blanckaert et al., 2017; Froman et al., 2018, 2019; Hurther & Thorne, 2011; Naqshband, Ribberink, Hurther, & Hulscher, 2014; Naqshband et al., 2017; Revil-Baudard et al., 2015, 2016; Stark et al., 2014; P. Thorne et al., 2018).

We have developed a multi-frequency coherent Doppler profiler (MFDop) for measuring vertical profiles of turbulence, stress, and sediment flux in the bottom boundary layer. In contrast to similar laboratory systems (for example the ACVP described by Hurther et al. [2011]), the MFDop has been developed specifically to make these measurements possible under field conditions by assuring a ruggedized design and robust system geometry (see Hay et al. [2014]). The design concept for this system was based upon experience...
The remaining challenge is to validate the measurements of bedload that can be made with bistatic Doppler systems and to understand the limits of the measurement. In particular, for the typical downward directed beam geometries of these systems shown in Figure 1, the backscattered signal from locations near the bottom are subject to interference from acoustic sidelobes. In addition, signal processing algorithms can have difficulties because of the large and rapid variations in signal levels that occur for ray paths that intersect the sediment-water interface. We approach the validation question via observations in a large-scale experimental flume where we compare the acoustic flux measurements with estimates made from bedform evolution, sediment traps, and predictions based on bed shear stress estimates. While the present work focuses on a particular bistatic system specifically designed for bottom boundary layer studies, the results are also relevant to measurements of bedload movement recorded with other Doppler profiling systems.

This paper is organized into three main sections followed by conclusions: SAFL Experiment describes the experiments undertaken in the St. Anthony Falls Laboratory (SAFL) and describes details of the SAFL main flume facility and bedload measurements that were made using: changes in migrating bedforms, direct measurements with the MFDop system, and the sediment traps built into the SAFL flume; Bedload Transport provides a comparison between the three transport measurements; and, Comparison with Predictions uses basic flow observations to make predictions of sediment transport based on estimated stress. Lastly, Discussion provides comments on some of the challenges with the experiment and measurements and Conclusion provides a brief summary of the results and concluding comments.

2. SAFL Experiment

2.1. SAFL Facility

The experiments were carried out in the Main Channel flume of the SAFL at the University of Minnesota. The flume is 1.8 m deep, 2.74 m wide and 80 m long and allows flow speeds of order 1 ms⁻¹ (Figure 2a). The water used in the flume is obtained directly from the immediately adjacent Mississippi River. Flow rate is established by adjusting the height of entrance and exit weirs. Bedload sediment transport is monitored continuously using five weigh pans spanning the full width of the flume (Marr et al., 2010). When each pan reaches a specified threshold weight, it automatically empties its captured load into a recirculation system which transports the sediment back to the head of the flume (see Figure 2a). The weight of sediment in each pan is sampled at 1 Hz and summed to obtain the bedload transport. The SAFL facility also includes a sonar mounted on a motor-driven cart riding on rails atop the flume side walls for bathymetric surveys. Additional measurements provided by the facility include water level and water volume discharge.

For the experiments presented here, a 40 cm thick bed of sand with a median diameter \(d_{50} = 0.40 \text{ mm} \) \( (d_{16} = 0.27 \text{ mm}, d_{84} = 0.63 \text{ mm}) \) was laid in the tank extending through the full 55 m length of the test section. Prior to beginning a set of trials at different discharges, the sand bed was raked flat by hand. The
flume was filled to a prescribed water level corresponding to \( \approx 1 \) m water depth, then left overnight with no water flow to allow air trapped in the sand bed to escape. In the morning, the discharge rate was adjusted through manipulation of the upstream and downstream weirs. The measurements reported here are from four trials at discharges from 1.80 to 2.20 m\(^3\)/s, corresponding to nominal flow speeds from 0.67 to 0.81 m/s\(^{-1}\) (see Table 1). Very little sediment movement was observed for discharges below about 1.50 m\(^3\)/s. A data acquisition sequence with the instrument array was initiated once the bedform (dune) field had become established, usually between 2 and 3 h after starting the flow. Data acquisition continued for the time

![Figure 2](image)

**Figure 2.** (a) Schematic of Saint Anthony Falls Laboratory’s main flume (taken from Marr et al. [2010]), the figure is not to scale. The 4.8 m section labeled Sonar zone corresponds to the region sampled for bedforms. Pressure sensors are located 5.88 and 38.80 m upstream of the acoustic instruments. \( \times \) marks the location of the acoustic instrumentation. (b) Details of acoustic instrumentation located at \( \times \). The relative placement of instrumentation is drawn to scale except for the along channel position of the sediment trap weigh pans.

| Trial | Discharge [m\(^3\)/s] | Velocity [m/s] | Stabilization time [h:min] | DAQ time [h:min] | Wavelength [M] | Amplitude [M] |
|-------|------------------------|----------------|----------------------------|------------------|----------------|---------------|
| 1     | 1.80                   | 0.67           | -                          | 4:47             | 0.80           | 0.05          |
| 2     | 1.96                   | 0.73           | 2:52                       | 3:36             | 1.5            | 0.09          |
| 3     | 2.04                   | 0.76           | 3:50                       | 3:07             | 1.8            | 0.13          |
| 4     | 2.20                   | 0.81           | 2:24                       | 1:55             | 2.2            | 0.17          |

Abbreviation: DAQ, data acquisition time.
taken for one complete dune to pass beneath the instrument array, typically 1–2 h. At the lowest discharge (1.80 m$^3$s$^{-1}$), no obvious large-scale bedforms developed so that the data acquisition period was somewhat arbitrary in that case (see Table 1).

2.2. Instrumentation

We deployed an array of instruments across the flume at a distance of 3.5 m upstream of the weighing pans. The instruments included the MF-Dop, a Nortek Vectrino, a Simrad Mesotech rotary pencil beam sonar, and two pressure sensors. Their positions within the SAFL flume are shown in Figure 2b.

A record of the bedform structures for any given flow condition was made using a cart mounted sonar (see Figure 2a). At the end of each data acquisition sequence, the flume discharge was stopped and a bed elevation survey was collected in five along-flume transects spanning a 5 m test section just upstream of the instrument frame. Each survey took $\lesssim$ 15 min to complete. The bedforms observed with the cart sonar during the four trials are shown in Figure 3. Over the 4 days of increasing flow speed, the observed bedform amplitudes increased from about 5 cm in trial 1 to about 20 cm in trial 4. The evolution of bedforms as they migrated past the acoustic instrumentation changed the water depth at the MF-Dop system from between 1.0 and 1.2 m as shown by the time series shown in Figure 4.

Figure 3. Results from SAFL sonar surveys of bedforms corresponding to trials 1 to 4 with discharges of 1.80, 1.96, 2.04, and 2.20 m$^3$s$^{-1}$ respectively. Observations for the four trials were made in sequence on April 29, May 1, May 2, and May 3 of 2015. SAFL, St. Anthony Falls Laboratory.

Figure 4. Depth of water as estimated by the MF-Dop mounting position and the range to bottom for trials 1–4 with discharges of 1.80, 1.96, 2.04, 2.20 m$^3$s$^{-1}$ panels a, b, c, and d respectively. Red, blue and green marks correspond to data collected using 1, 3, and 5.25 $\mu$s transmit pulses. The axes scales are the same for all trials. MF-Dop, multi-frequency coherent Doppler profiler.
2.3. Multi-Frequency Coherent Doppler Profiler

The MFDop uses a bistatic geometry to allow three component velocity measurements in a near vertical profile; the basic instrument geometry is shown in Figure 1. The central active transducer 1 and passive receiving transducers 2 and 3 have identical beam patterns and are arranged in isosceles triangle geometry. This geometry is used because, in the case of an ideal, uniform scattering domain, the resulting measured velocity component would be along the line bisecting the two transducers, independent of range (Hay et al., 2012) (an example of a single bistatic beam pattern is shown in Figure 5). Measurement of horizontal velocity requires the combination of component measurements from two pairs of the symmetrical transducers (for example one velocity from transducers 1 and 2 and a second velocity from transducers 1 and 3 as shown in Figure 1). The nominal sample location for the resulting velocity estimate would lie along the axis of transducer 1, but the measurement includes contributions separated by as much as 4 cm at the extreme limits of the measurement profile. These offsets will lead to some loss in resolution of turbulent velocity components as discussed by Rolland (1994). The use of receiver transducers with wide beam-widths such as described by Hurther et al. (2011) would reduce these offsets; we have not used this approach because of the associated loss in system sensitivity that results with less focused acoustic beams.
An example of the sampling space for the MFDop system is constructed from transducer beam patterns and backscatter profiles expected over a mobile sediment bottom (Figure 5). For the MFDop system, optimal measurements can be acquired over an interval from 65 to 95 cm below the instrument because in this region, the 3 dB sample width (indicated by green dots in Figure 5) is constrained to within ± 1 cm of the beam bisector (blue crosses in Figure 5). Velocity measurements are possible outside this range, but they are characterized by much greater uncertainties. For much of the time during the laboratory trials, the bottom occurred at a range of about 70 cm and for this reason, we have matched that position in Figure 5. Figure 5 shows that the sample points overlay the target points to within 2 mm for \(z\)-positions greater than −55 cm (a height above the bottom of about 15 cm). Alignment is within 1 cm between \(z = −45\) cm and −55 cm but it is critical to note that the width of the sample domain becomes progressively wider expanding to about 10 cm at \(z = −45\) cm. At even greater heights, the sample points no longer agree with the beam bisector line and the data cannot be considered of any use. In the present work, we were interested in the sampling just above the bottom so that these deviations at higher levels were of no concern.

The MFDop profiles were acquired in 20-ping ensemble averages. Three different transmit frequency configurations were used: single, dual, and four frequency (see Table 2). Single-frequency transmission allows the maximum bandwidth of the system to be used, providing maximum range resolution, but does not adequately constrain the inversion of backscatter data. Data were collected in time-series of about 2.5 min duration; slight differences in timing arose because of ping rate adjustments used to minimize interference from surface reflections.

### 2.4. Pencil Beam Measurements

A rotary pencil beam sonar was used to collect along-flume bed elevation profiles centered at the MFDop position (see Figure 2b). These nominally 5-m long profiles were acquired in triplicate and ensemble-averaged immediately following every set of three MFDop data collections, that is, approximately every 10 min. The ensemble-averaged profiles were processed to detect the crests and troughs of bedforms present in each profile, subject to a minimum bedform height of 1 cm and minimum crest separation of 30 cm. Cases in which the same crest-to-trough feature appeared in successive bed elevation profiles were then identified, and used to obtain geometric estimates of bedload transport as illustrated in Figure 6. Two methods were used to calculate the transport. The first method calculated the dune-averaged transport per unit cross-stream distance using the widely used dune-tracking method (see Claude et al. [2012] and references therein),

\[
Q_x = (1 - \rho)c_ds_d/L_d,
\]

where \(\rho\) is the bed porosity, \(c_d\) is the dune celerity (estimated based on a lagged-correlation between two successive profiles), \(S_d\) is the longitudinal cross-sectional area of the dune, and \(L_d\) is the dune wavelength. In the second method, the area of the shaded region indicated in Figure 6 was used to represent the sediment accumulated on the downstream face of the dune during the time between two successive profiles. Note that the second method is a measure of transport over the dune crest, which is
expected to be larger than the dune-averaged transport measured by the first method. For both methods, conversion from bed area per unit time to transport (sediment volume per unit time per unit stream width) assumed a sediment density \( \rho_s = 2650 \text{ kg/m}^3 \) and porosity \( p = 0.40 \). Finally, results from each method were averaged over all of the individual dunes detected in a given bed elevation profile pair, typically 1 to 2 dunes.

### 2.5. MFDop Sediment Flux Measurements

MFDop data collected using the 4-frequency configuration were converted to sediment grain size and concentration following a statistical method (see Wilson & Hay, 2015), with MFDop system sensitivity calibrated using the standard target technique described in Wilson and Hay (2017), and sediment scattering properties following Moate and Thorne (2013). The 20-ping ensemble-averaged backscatter measurements imply a theoretical noise level of \( \pm M / \sqrt{20} \), or 20\%, where \( M \) is the measured concentration. Direct validation was not possible, due to an inability to obtain independent measurements of range- and time-resolved concentration profiles for comparison. Previous measurements with this technique (e.g., Wilson & Hay, 2015) have shown that the time-averaged measurement is accurate, although biases may occur if there are errors in the assumed backscatter form factor; the results by Moate and Thorne (2013) suggest such biases may be of order 10\%. The inversion method will break down at higher concentrations but, accurate results have been demonstrated at concentrations of order 10 kg/m\(^3\) (Wilson & Hay, 2015) and, Hurther et al. (2011) and Thorne et al. (2018) report stable results at concentrations of order 100 kg/m\(^3\).

In the inversion process, acoustic attenuation in the water column consisting of a background uniform attenuation in water, and a concentration-dependent attenuation due to sediments were included. The former, attenuation in water, was estimated empirically based on an exponential fit to the depth-dependent time-averaged (over each data acquisition) backscatter intensity. This fit was confined to the uppermost 20 MFDop range bins, where no suspended sediment was present and where the observations were well-fit to the exponential model which implies depth-uniform attenuation. Notably, the results yielded attenuation ranging from 0.4 to 0.6 Np m\(^{-1}\), much larger than that expected for clear water alone, possibly due to an abundance of microbubbles and/or fine wash-load sediments present in the natural river water used in the flume. Indeed, the water running through the flume was visibly turbid, and bubbles were omni-present likely because of the waterfall used to draw water from the Mississippi River. The fitted data were therefore used to correct the full MFDop profiles, assuming uniform contributions to attenuation and backscatter intensity; this correction resulted in 3–6 times larger MFDop-derived sediment concentrations at 75 cm range.

An example of data collected with the MFDop system, at a discharge of 1.96 m\(^3\)s\(^{-1}\), is presented in Figure 7. The x-component of velocity (Figure 7a) falls to zero at the bottom but with small non-zero velocity values beyond the bottom. Sediment flux estimates are created by multiplying velocity with concentration and, the resulting sediment transport is dominated by the bedload near the bottom (see Figure 7c). The location of the bottom (at height = 0 cm in Figure 7) has been identified as the range at which the vertical gradient of the backscatter intensity is maximal. It is most likely that this location occurs somewhere in the bedload layer rather than the stationary bottom. Profiles of sediment concentration (Figure 7b) show concentration decreasing with depth beyond the peak location; this decrease results from the breakdown in the inversion algorithm as progressively less signal is received from greater depths within the sediment.

The peak in sediment flux occurs in the near bottom region where the velocity is very small compared to values higher up in the water column. A structure of particular concern was the occurrence of negative flux in regions beyond the bottom; this structure was prevalent in measurements made during the 2.04 m\(^3\)s\(^{-1}\) discharge observations. That particular trial stood out in that the bed level was consistently high (consider Figure 4c) placing the observations near the top of the MFDop operating interval. Horizontal components of velocity are extracted by forming differences between the independently measured beam components; when these sample locations are spatially separated as is caused by the side lobe structures, the velocity inversion becomes inaccurate. Sidelobe scatter from beyond the bottom location yielded negative velocities at this time, which gives rise to the interval of negative transports seen beyond 1.5 cm below the bottom (see Figure 7c). In order to avoid these negative contributions, the transport was integrated from 0.5 cm above the bottom to 1.5 cm below the bottom. The choice of the upper level is not critical as the flux rapidly goes to
zero above the bedload. The lower level was selected because we frequently observed (unrealistic) negative velocities beyond about \( z = -1.5 \) cm. The negative velocity at about 1 cm height in Figure 7a is likely the result of a small misalignment between complementary acoustic beams in the presence of the large velocity gradient in that region. Spurious velocities in this region did not influence sediment transport estimates because the suspended sediment concentration is comparatively low at this level. Velocity gradients in the bedload are much smaller than in the transition between water and sediment velocities reducing the effects of any misalignment.

We note that the transport profile in Figure 7c indicates a bedload layer thickness of about 2–3 cm which should be adequately resolved by the \( \sim 0.4 \) cm depth bins.

### 3. Bedload Transport

Bedload transport in the present experiment has been estimated in three independent ways: by depth integrating the MF Dop flux profiles around the detected bottom depth, by using the sediment trap weigh pans built into the SAFL flume, and by estimating sand dune transport with the pencil beam data. These observations are shown as time series in Figure 8. For the sequence of four experiments (spread over a 5-day interval), the discharge was increased from 1.80 to 2.20 m\(^3\)s\(^{-1}\). Pencil beam transport estimates were not available for the lowest discharge because of the absence of well-formed dunes. All of the measurements indicate a trend toward increasing transport with increasing discharge, but they also all exhibit significant variability.

Figure 7. Example of flow speed, concentration, and sediment flux measurements made with the MF Dop system over one 2.5 min sampling interval during a discharge of 1.96 m\(^3\)s\(^{-1}\). Dashed green lines indicate the region integrated to determine bedload transport.
Figure 8. Time series of observed sediment transport: black line indicates weigh pan measurements, blue shows MFDop calibrated observations, and red shows transport inferred from pencil beam data. Horizontal bars indicate the time interval during which the bedforms were considered to have equilibrated to the flow conditions.

demonstrating the temporal and spatial variability of the transport itself. Temporal variability was caused by fluctuations in the flow field, as will be shown later in Section 4.1 when comparing to bedload transport predictions. Spatial variability was dominated by the large migrating dunes that by their nature lead to significant changes in bedload as they travel past any fixed point. These dunes were not straight-crested (see Figure 3) so that significant differences in measurement would be expected depending on the cross-flume observation point. As the experiment was configured, the pencil beam and MFDop observations were closely spaced in the cross-channel direction, but the MFDop is a point measurement while the pencil beam estimates are spatially integrated. Similarly, the weigh pan measurements are spatially averaged in the sense that they span the width of the flume. Comparisons were also made using only weigh pan 3—the pan in line with the MFDop and pencil beam—but those data (not shown) actually increased the variability and so comparisons here make use of the overall cross-channel average.

The measurement variability can be reduced by time averaging over observations at a particular discharge. Such averaged data are shown in Figure 9. The MFDop observations agree within uncertainty with SAFL weigh pan data at the two higher discharge rates but not at the lower two. The pencil beam data based on the whole dune motion are consistent with the weigh pans (although they do not agree with experimental uncertainty). The transport estimates based on dune faces alone are high by perhaps a factor of 2, a result that is consistent with such selective sampling over the bedforms.

4. Comparison with Predictions

Bedload transport can be estimated using empirical relations (see discussion in Appendix); we have used predictions based on work by Meyer-Peter and Müller (1948), and Nielsen (1992) through Equations A.2 and A.4 respectively to predict bedload for the range of conditions corresponding to the SAFL experiments. Both equations require estimates of the Shields parameter, which combines sediment characteristics with flow conditions expressed through the friction velocity, \( u_* \). The shear stress was estimated from the along-channel momentum balance. The data from the two pressure sensors mounted on the flume sidewall above the level of the sediment and separated by 33 m (see Figure 2a) were corrected for atmospheric pressure using the hourly values recorded at the Minneapolis airport. These data provided both the along-channel pressure gradient and the mean water depth \( \psi \), after accounting for the height of the pressure sensor above the mean bed level, as determined from the cart sonar surveys.

4.1. Bedload Transport Predictions

The single point velocity measurements from the Nortek Vectrino (\( U_c \)) were used to determine the vertically averaged flow speed (\( \overline{U} \)) using the log law relations presented in Equations A.8 and A.9. It was found that \( \overline{U} \sim 0.9 U_c \), where the difference is not large as might be expected since the Vectrino sample was located far from the bed; i.e., the Vectrino height, \( z_V \), was between 0.65 and 0.75 m, and water depth \( \psi \sim 0.9\text{–}1.0 \) m. Values of \( u_* \) and \( \overline{U} \) were used to create estimates of drag coefficient through Equation A.7, these values are shown in Figure 10a. The estimates of \( u_* \)
were adjusted to skin friction values \( u' \), using Equation A.10, and values of the skin friction Shields parameter shown in Figure 10b were determined using Equation A.3. Finally, values of bedload transport \( (Q_B) \) from the Meyer-Peter and Müller (MPM) relation, Equation A.2 and Nielsen’s relation (A.4) are shown in Figure 10c.

The predicted transport values are similar to the direct observations (Figure 8), with values increasing with increasing discharge and again, there is significant scatter at any one discharge. Note that the scatter in these predictions was almost entirely due to variations in the flow speed, which increased with increasing flow rate to a maximum of order 10% fluctuations (standard deviation as a percentage of mean) in \( \bar{U} \) and \( u' \) at the highest flow rates measured (see Figure 11). The predictions based on Nielsen’s relation, Equation A.4, are consistently higher than those using the Meyer-Peter and Müller (MPM) relation, Eq. A.2. The difference is more obvious for higher flow conditions, as expected (Nielsen, 1992).

The predicted values in Figure 10c were averaged over the periods for which measurements were taken at a given discharge. The resulting values are plotted in Figure 9. The MPM values are very close to the SAFL weigh pan measurements. In contrast, the values obtained using Nielsen's relation are higher than the weigh pan measurements (by about a factor of two). The MFDop data track the MPM predictions closely. It must be stressed that the theoretical predictions are entirely independent of the MFDop measurements.

5. Discussion
5.1. MFDop Measurements

For the MFDop system, sediment transport estimates require both velocity and sediment concentration profiles. Sediment concentration was extracted from the single vertically directed acoustic beam using the inversion technique presented by Wilson and Hay (2015). One of the difficulties in establishing these inversions was the high levels of scattering that were observed throughout the water column. The scattering appears to have been due to entrained bubbles and/or high concentrations of fine sediment in suspension and was well-fit using an assumption of a uniform background absorption. Whatever the source, observed acoustic absorption rates were 0.4–0.6 Np m\(^{-1}\) and these elevated absorption rates required a correction to sediment concentration by a factor of between 3-6 at 75 cm range as described in Section 2.5. Transport rates for the MFDop were estimated by multiplying velocity profiles by concentration profiles and then integrating over depth; the results are necessarily dependent on the accuracy of the concentration measurements. We expect concentration measurements to have an accuracy of about 30% but the inversion technique will eventually fail at high sediment concentrations due to the effects of multiple scattering. Hurter and Thorne (2011) report maximum concentrations of order 100 kg/m\(^3\) in bedload transport and that value is comparable to the maximum observed values in our trials. While we have no direct verification of accuracy at high concentrations, the fact that MFDop measurements agree with independent transport measurements here (and as reported in Hurter & Thorne [2011]) suggests that movement of material at concentrations greater than this measurement limit did not contribute to bedload transport.

Velocity measurements near the bottom are more complicated than those of concentration because they require combinations of data from multiple, bistatic, intersecting acoustic beams. Near the bed, the highly reflective sediment-water interface results in non-negligible contributions from beam pattern sidelobes to the backscatter amplitude, and so in this region, the sample points deviate from the bisector line (see Figure 5).
The degree of deviation depends on the transducer beam patterns, the location of the bottom, and the length of the sample bins in relation to the bedload thickness. But, it must be stressed that this deviation in sample locations is a characteristic of all bistatic systems and is not specific to the MFDop geometry. A model of acoustic backscatter (Zedel et al., 2015) was used to explore the beam pattern characteristics for the MFDop system sampling above a bottom with a simulated suspended sediment and bedload profile (Wang & Yu, 2007). An example of the interplay between pulse-length, bedload thickness, and sample position is indicated by the bistatic beam maps shown in Figure 12.

Figures 12a and 12c present simulations of high resolution sampling for the MFDop system with a bedload layer of thickness 0.3 and 1.0 cm respectively. The strong backscatter gradient in the bedload region leads to offsets in the actual sample location. Within the layer and for about 2–3 mm above the layer, the sample location is biased toward more positive x-positions which will tend to decrease the magnitude of the measured velocity component. A positive offset of 1 cm (the largest indicated in Figure 12c) will bias (horizontal) velocity estimates low by about 10% in this location. Toward the bottom of the bedload layer, the sample locations deviate toward negative positions which become comparatively large (offsets in excess of 3 cm are seen in the example simulations shown). This transition to negative offsets is expected as contributions from sidelobes on the positive side of the sample location are no longer received. These simulations likely exaggerate that negative bias as the model completely extinguishes signals that would be received from within the stationary bed. The same sequence of sample offsets is seen when using larger sample intervals (Figures 12b and 12d) but the magnitude of the offset is reduced because of the spatial averaging. While not strictly quantified, we expect that errors due to offsets in the sample location will lead to errors in mean velocity estimates of about 10%.

A separate problem associated with offsets in sample location occurs when there is a lack of symmetry between offsets in complimentary beam pairs as will happen in the presence of a sloping bottom. We expect that this process has led to the occurrence of unphysical negative velocities in our MFDop data. In order to suppress this contamination, the integration to determine transport was terminated 1.5 cm beyond the detected bottom. The choice of this cutoff was based on inspection of the signal level and velocity profiles and the occurrence of velocities that were judged to be spurious as they were directed opposite to the flow velocity just above the bottom. Given that the spurious velocities are associated with sidelobe effects, changes to system geometry and operating configuration would require adjustment to this chosen cutoff value.

Error in the present study is also introduced because of unresolved velocity scales associated with the spatial separation of velocity components that are combined to make individual velocity measurements. When these velocities are combined with concentrations to estimate sediment flux, there is further potential for the introduction of bias. Turbulent transport arises when velocity and concentration fluctuations are correlated. Offsets in the sample locations for the combined measurements will alter the degree of correlation, most likely reducing that correlation and underestimating the contribution to transport. In a similar flume-based sand dune experiment, Naqshband, Ribberink, Hurther, Barraud, et al. (2014b) found that the turbulent sediment flux can represent as much as 50% of the total bedload transport. Given the spatial averaging and sample offsets that occur with the MFDop system, the affected (horizontal) length scales are those that are less than 3 or 4 cm. We note that this bias can be reduced in bistatic systems by focusing the velocity sample points using omnidirectional receivers (as discussed by Rolland [1994]).
The MFDop transport estimates were compared to independent estimates derived from the dune migration rates and the sediment traps in the SAFL facility. The MFDop measurements and the estimates based on whole dune migration rates agreed very well with the SAFL weigh pan observations. Estimates based on the migration of the dune face alone yielded values that were consistently high by about a factor of 2.

All of the measurements showed substantial variability in time which was reduced by averaging over periods of ~2 h or more. The MFDop measurements showed greater variability than the dune-based measurements but this is perhaps expected as the MFDop measurement is strictly based on a single profile location while the dune measurement is based on data averaged over an entire dune structure. The bulk measurements of transport (weigh pans and dune migration) implicitly include all turbulent components of transport. The MFDop measurements will also include turbulent components but these are limited (primarily) by the spatial resolution of the system. What the present results do indicate is that the contributions of turbulent transport that are not being resolved by the MFDop system are small compared to the overall uncertainties in the comparison.

**Figure 12.** Bistatic beam pattern in the interval 2 cm above the bottom located 70 cm below the MFDop instrument. Blue crosses identify the beam bisector, red crosses identify the realized sample location, green dots identify the half-power point of the samples and white lines indicate the travel time bin boundary for a single sample bin: (a) sample locations with 1 μs transmit pulse, 0.3 cm thick bedload, and (b) 5.25 μs transmit pulse, 0.3 cm thick bedload, (c) 1 μs transmit pulse, 1.0 cm thick bedload, and (c) 5.25 μs transmit pulse, 1.0 cm thick bedload. MFDop, multi-frequency Doppler profiler.
6. Conclusions

We have reported on measurements of bedload transport using a wide bandwidth, pulse coherent Doppler sonar: the MFDop system. MFDop bedload transport estimates made by multiplying Doppler velocity measurements with concentration from acoustic backscatter were consistent with the weigh-pan estimated transports, and with estimates made from bedform migration rates (based on the whole dune structure). Predictions by the MPM relations were very close in agreement with the direct MFDop observations and transport extracted from the (whole) dune migration rate. In contrast, the predictions using Nielsen’s relations were generally higher than our observations (by about a factor of two or more) and were more consistent with transports based on the dune faces alone.

There are difficulties when making velocity measurements in the bedload region using acoustic systems with bistatic configurations such as used by the MFDop system. We estimate that offsets in sample locations caused by acoustic sidelobes interacting with the bottom are likely to introduce velocity errors in the order of 10% in the near-bed and bedload regions. While the exact size of these errors is specific to the MFDop configuration, these shifts in sample geometry are a fundamental characteristic of all bistatic systems. A more significant error source in estimating bedload transport is likely the inversion of acoustic backscatter levels to estimate sediment concentration; we expect that these inversions are typically accurate to within 30% in the present study. The inversion algorithm breaks down as signal levels decrease due to acoustic scattering and absorption within the high-concentration bedload layer. The maximum concentrations we observed in the bedload regions were in the order of 100 kg/m$^3$ (consistent with previous studies Hurther et al. [2011]; Thorne et al. [2018]). This sediment concentration is still far below the 1,650 kg/m$^3$ expected of the non-moving bottom so that some contribution to the bedload transport is being missed.

Acknowledging that there are challenges when using Doppler systems in the presence of a strong scattering gradient, the present results nevertheless demonstrate that accurate measurement of bedload transport are possible. These Doppler techniques then provide the potential of non-invasive and large-scale surveying of sediment transport in rivers and coastal areas. Other observations are invasive or based on aggregate averages over some region or time and can provide no insight into the detailed spatial and temporal characteristics of the transport process.

Appendix A: Bedload Transport Prediction

The total sediment transport is normally separated into bedload and suspended load components (Sleath, 1984). Bagnold (1956) defines bedload as that part of the total load which is supported by inter-granular forces rather than a fluid drag; in practice, however, the definition of bedload is often more arbitrary, for example, the part of the total load which is captured by sediment traps (Nielsen, 1992). The volume rate of bedload transport can be predicted semi-empirically by Meyer-Peter and Müller (1948),

$$Q_b \propto (\tau - \tau_c)^\xi$$  \hspace{1cm} (A.1)

where $Q_b$ is the volume rate of sediment transported as bedload, $\tau$ is the shear stress at the bed, and $\tau_c$ is the critical shear stress for sediment motion (Shields, 1936). The exponent $\xi$ is often taken to be 3/2, but values as large as 7/2 have been suggested (Dyer & Soulsby, 1988). Laboratory results (Ribberink, 1998) yield a value of 1.67 that is closer to 3/2.

When stress is expressed in terms of the Shields parameter, $\theta$, $Q_b$ can be expressed as:

$$Q_b = 8B(\theta' - \theta_c)^{3/2}$$  \hspace{1cm} (A.2)

where $B = d_{50}\sqrt{(s - 1)gd_{50}}$, $d_{50}$ the median grain diameter, $s$ the specific gravity of the sediment grains, $g$ the acceleration due to gravity, $\theta_c$ is the critical Shields parameter, set to 0.05 here (Nielsen, 1992), and $\theta'$ is the skin friction Shields parameter, given by

$$\theta' = \frac{u_*^2}{(s - 1)gd_{50}}$$  \hspace{1cm} (A.3)
with \( u^* \) the friction velocity associated with skin friction. Nielsen (1992) proposed a slightly modified version:

\[
Q_b = 12B\sqrt{\bar{\theta}} (\bar{\theta}' - \theta_0).
\]  
(A.4)

In general, Nielsen (1992)'s formula is expected to be more accurate for \( \bar{\theta}' \gtrsim 0.3 \).

### A.1 Drag coefficients

The friction velocity \( u^* \) for steady unidirectional flow can be estimated from the vertically integrated momentum equation for steady flow assuming a balance between horizontal stress and the pressure gradient terms, yielding

\[
u^2 = -g\psi \eta_x
\]  
(A.5)

where \( \eta \) is the surface elevation, and \( \psi \) is the mean water depth. The subscript \( x \) denotes partial differentiation. Letting \( U_v \) represent the mean flow speed registered by the Vectrino at a height \( z_v \) above bottom, the drag coefficient referenced to \( z_v \) is

\[
C_{dv} = u^2 / U_v^2
\]  
(A.6)

The drag coefficient referenced to the vertically averaged flow speed, \( U \), is given by

\[
C_{dv} = u^2 / \bar{U}^2
\]  
(A.7)

In order to estimate \( \bar{U} \), the log law (Monin & Yaglom, 1971)

\[
U(z) = \frac{u_* \ln z}{\kappa} \ln \frac{z}{z_0}
\]  
(A.8)

was invoked, with the von Kármán constant \( \kappa = 0.4 \), yielding

\[
\bar{U} = \frac{1}{\psi} \int_{z_0}^{z_v} U(z)dz
\]  
(A.9)

with the lower limit of integration given by the roughness parameter, \( \ln z_0 = \ln z_v - \kappa U_v / u_* \).

### A.2 Skin Friction

Bedload is driven by skin friction, whereas \( u_* \) estimated from the surface slope includes contributions from both skin friction and form drag. In order to estimate the skin friction contribution alone, the method developed by Einstein (1950) for flow over dunes is implemented. The method involves finding the value of \( u_*' \) satisfying

\[
\bar{U} / u_* = 6 + 2.5 \ln (\delta_i / k_s)
\]  
(A.10)

where

\[
\delta_i = u_*^2 / gI,
\]  
(A.11)

\( I = u_*^2 / g\psi \) being the so-called energy slope, \( k_s = 2.5d_{50} \) (Nielsen, 1992), the equivalent Nikuradse grain roughness. The value of \( u_* \) estimated from Equation A.10 enables estimates of bedload transport to be made using Equations A.2 and A.4.

### Data Availability Statement

Zedel et al. (2020) provided online access to data and MATLAB scripts used in the preparation of this paper.
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