ON THE INFLUENCE OF GRAVEL BED DYNAMICS ON VELOCITY POWER SPECTRA

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On the influence of gravel bed dynamics on velocity power spectra

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A series of flume experiments were conducted to study the effect of bed-form dynamics on the flow over a gravel bed comprised of a wide grain size distribution. Spectral analysis of the measured velocity fluctuations reveals two distinct scaling regimes separated by a spectral gap. The relatively high frequency range of scales is dominated by turbulence and characterized by a clear Kolmogorov spectral scaling. A second lower-frequency scaling range is the signature of large-scale velocity changes associated with the relatively slow evolution of the bed topography.

1. Introduction

Measurement of turbulent flow structures in a gravel-bedded environment has received considerable attention in the past few decades; yet, there is still debate about the origin and development of these flow structures and, in turn, their influence on the bed surface itself [*Nelson et al.*, 1993; *Nelson et al.*, 1995; *Nikora et al.*, 2002; *Schmeeckle et al.*, 2007]. It has been suggested in the literature that the initiation of gravel movement is strongly influenced by transient flow structures with time scales of about 1-10 seconds which are superimposed on the more random turbulent flow field [*Drake et al.*, 1988; *Kirkbride and Mclelland*, 1994; *Kirkbride and Fergusson*, 1995]. Over a rough boundary, such as in a gravel-bedded channel, friction created by individual gravel particles or clusters of particles (i.e., microtopography as well as bedforms) retards the flow velocity, but the effect diminishes with increasing height above the bed. This roughness creates near-bed turbulence which is responsible for entrainment of particles predominately linked to sweeps, bursts and larger coherent structures [*Best*, 1993; *Jerolmack and Mohrig*, 2005].

In this paper we use long time series of bed elevation and velocity fluctuations measured over a gravel-bedded channel to quantify both flow structures and bed structures. The results clearly demonstrate the signature of bed structures on the near-bed velocity fluctuations. The paper is outlined as follows. In section 2 a brief description of the experimental setup, the data analysis techniques and the results is presented. In section 3 the discussion and interpretation of the results are given, followed by the conclusions in section 4.

2. Analysis of experimental data

2.1. Experimental Setup

Experiments were conducted in the Main Channel facility located at St. Anthony Falls Laboratory, University of Minnesota. The channel is 55 m long, 2.74 m wide and has a maximum depth of 1.8 m with a maximum discharge capacity 8000 l/s . It is a partially sediment recirculating channel while the water flows through the channel without recirculation. Intake of the water in the channel was directly from the Mississippi river. The bed of the channel was composed of a mixture of gravel and sand with a median particle size diameter, $d_{50} = 7.7$ mm, $d_{16} = 2.2$ mm and $d_{84} = 21.2$ mm. A priori to data collection a constant water discharge, Q, was fed into the channel to achieve quasi-dynamic equilibrium in transport and slope adjustment for both water surface and bed. After attaining the equilibrium, experiments were ran for approximately 20 hrs. (More details about the experimental setup can be found in *Singh et al.* [2008, 2009]).

The data reported here are the velocity fluctuations (in the flow direction) and the temporal bed elevation collected at the downstream end, along the centerline of the channel. The velocity fluctuations were measured using Acoustic Doppler Velocitimeter (ADV) at an approximate distance of 10 cm above the mean bed level. For the bed elevation fluctuations submersible sonar transducers of 2.5 cm diameter were deployed 0.3 m (on an average) below the water surface. The sampling interval of bed elevation measurements was 5 sec with a vertical precision of 1 mm. Figure 1 shows the setup of ADV and the Sonar placed at the downstream end of the channel. Measurements were taken over a range of discharges corresponding to different bed shear stresses. Here we report the data collected at discharges of 2000 l/s and 2300 l/s. Figures 2 and 5 show the time series of velocity fluctuations (top) and the corresponding bed elevation (bottom) for discharges of 2000 l/s and 2300 l/s, respectively. As the data was collected in the fall season, there were some leaves floating in the channel which might have resulted in spikes in the velocity and bed elevation data. Even though the amount of spurious spikes in the data was found to be very small, these were removed as part of the data treatment for erroneous measurements.

2.2. Fourier analysis of bed elevation and velocity fluctuation

Power spectral density (hereafter PSD) is a commonly used tool to measure the distribution of energy (variance) in the signal across frequencies (or scales). In other words, it shows at which scales the variations are strong and at which scales variations are weak. A simple way to estimate PSD is by taking the fast Fourier transform (FFT) of the signal [*Stoica and Moses*, 1997]. For a discrete signal $X(t)$, the power spectral density is given by

$$\Phi(\omega) = \left| \frac{1}{\sqrt{2\pi}} \sum_{n=-\infty}^{\infty} X(t) e^{-i\omega t} \right|^2 = \frac{\hat{X}(\omega)\hat{X}^*(\omega)}{2\pi} \tag{1}$$

where $\hat{X}(\omega)$ is the discrete Fourier transform of $X(t)$, $\hat{X}^*(\omega)$ is its complex conjugate and ω is the frequency. In our case the signal $X(t)$ is the flow velocity or the bed elevation. Special emphasis is placed here on identifying spectral scaling ranges, i.e., ranges of scales over which log-log linearity is observed in the power spectral density.

2.3. Results

The power spectrum of the velocity fluctuations (measured at 25 Hz) at a discharge of 2000 l/s is shown in Figure 3. Two clear scaling ranges can be observed, separated by a spectral gap. For relatively small scales (high frequencies) in the range of 0.1 sec to 2 sec, the slope of the PSD is $\sim$ -5/3, which corresponds to the Kolmogorov spectrum of turbulence. A second scaling range is observed for scales between 100 sec and 1.0 hour, for which the slope of PSD is $\sim$ -1.2. Figure 4 shows the PSD of the bed elevation (measured at sampling intervals of 5 sec). A clear scaling is also found in the elevation field, with a PSD slope of $\sim$ -2.0 for the scales of 13 sec to 45 min. Figures 6 and 7 show the power spectral density of the velocity fluctuation (measured at 200 Hz in this case) and the bed elevation (measured at 0.2 Hz) respectively for a discharge of 2300 l/s. The same spectral scaling ranges discussed above are found for the two flow conditions (discharges) considered here.

3. Discussion

Velocity fluctuations in turbulent flows have been previously analyzed in terms of their scaling properties, intermittency and characteristic shape of their PDF at different scales. Their PDF roughly evolves from a Gaussian shape near the integral scale to a stretched exponential shape near the Kolmogorov scale [*Malecot et al.*, 2000]. In the case of turbulent channel flow over topography, the statistics of the velocity field is affected by the roughness elements (bedforms) and their spatial distribution along the channel reach [*Robert et al.*, 1992, 1993; *Lamarre and Roy*, 2005]. For instance, the mean (time-averaged) velocity measured at some location over the crest of the dunes is larger than the velocity measured at the same elevation over the dune troughs. In addition, the PDF of the velocity fluctuation is found to be more skewed over the dune crest as compared with the dune trough [*Hassan and Reid*, 1990; *Best*, 1993].

In the hypothetical case of stationary bed forms, one would expect the power spectra of velocity measured at a given location to be affected solely by the turbulence and, consequently, characterized by a -5/3 scaling in the inertial subrange [*Kolmogorov*, 1961]. The evolution of the bedforms, however, introduces additional variability in the velocity field at the range of temporal scales associated with that evolution. This effect explains the existence of the second scaling range (between 100 sec and 1.0 hr) in the power spectrum of velocity, as shown in Figures 3 and 6. Notice that the largest scale in that range ($\sim$ 1.0 hr) corresponds with the integral scale of the measured bed elevation field presented in Figures 4 and 7. This largest scale of 45-60 min can also be seen in the time series of bed elevation (Figure 2 (bottom)) and is the characteristic time scale with which the largest bedforms move. The clear signature of the large-scale bedforms on the velocity time series and PDF suggests the potential of using relatively low frequency velocity measurements near the bed to detect the characteristic time scales associated with the evolution of bed topography. The spectral analysis of our velocity measurements also shows that this scaling range is separated from the turbulence range by a spectral gap, i.e., a range of scales with virtually no contribution to the velocity variance.

Bed elevation fields and their evolution are found to share important similarities with other natural surfaces such as landscapes. Landscapes present multiscale self-similar properties through a wide range of scales (see *Vening Meinesz* [1951]; *Newman and Turcotte* [1990]; *Pelletier* [1999]; *Passalacqua et al.* [2006] and references therein). In fact, *Passalacqua et al.* [2006] have documented that landscapes also share important similarities with turbulence since both systems exhibit scale invariance (self-similarity) over a wide range of scales and their behavior can be described using comparable dynamic equations. This similarity can be seen, for example, in the behavior of power spectra of the landscapes which exhibit a log-log scaling range with a slope of -2. Here also, we observe a slope of $\sim$ -2 in power spectra of bed elevation for both the discharges of 2000 l/s and 2300 l/s (Figure 4 and 7). Furthermore, *Singh et al.* [2009] have shown the multiscale behavior of the bed elevation (bed topography) for different flow conditions in a gravel-bedded environment. In that particular study they have quantified the slope of the second moment of the structure functions (which is related to slope of the PSD with a relation $\beta = 2H + 1$, where β is the slope of PSD, and H is the Hurst exponent), and found that it is similar to the slope obtained here in the PSD of bed elevation fluctuations (Figure 4 and Figure 7).

Two major priorities for further research are suggested by this work. First, what controls the slope of $\sim$ -1.2 for low frequency velocity fluctuations, and is this slope universal and independent of the flow? Second, does the length scale of the spectral gap depend on the flow discharge, depth wise position of velocity measurements in the channel or the grain size distribution of the bed material? For example, the range of scales in the velocity spectrum that are affected by the bedform evolution is likely to grow (and the spectral gap shrink) with decreasing distance of the velocity measurement with respect to the bed. In order to test this hypothesis and answer the above questions, future work will be focused on the behavior of velocity spectra measured at different positions along the depth of the flow as a function of varying discharge and grain size distribution.

4. Conclusion

This paper investigates the behavior of power spectral density of flow velocity and bed elevation time series measured in an experimental channel under two flow regimes. The power spectral density of the velocity shows two distinct power-law scaling ranges: a high-frequency range associated with turbulent eddy motions, and a low-frequency regime that represents the effect of the evolving multiscale bed topography on the velocity fluctuations at the measurement location. The two regimes are separated by an spectral gap.

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Figure 1. Experimental channel facility at the St. Anthony Falls Laboratory, University of Minnesota.

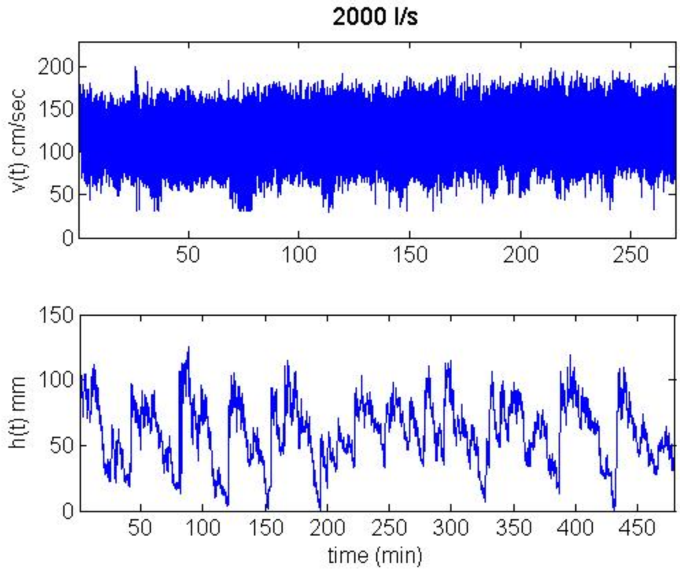

Figure 2. Velocity fluctuations measured at the downstream end of the channel at a frequency of 25 Hz (top) and the corresponding bed elevation measured at a frequency of 0.2 Hz (bottom) for a flow discharge of 2000 l/s.

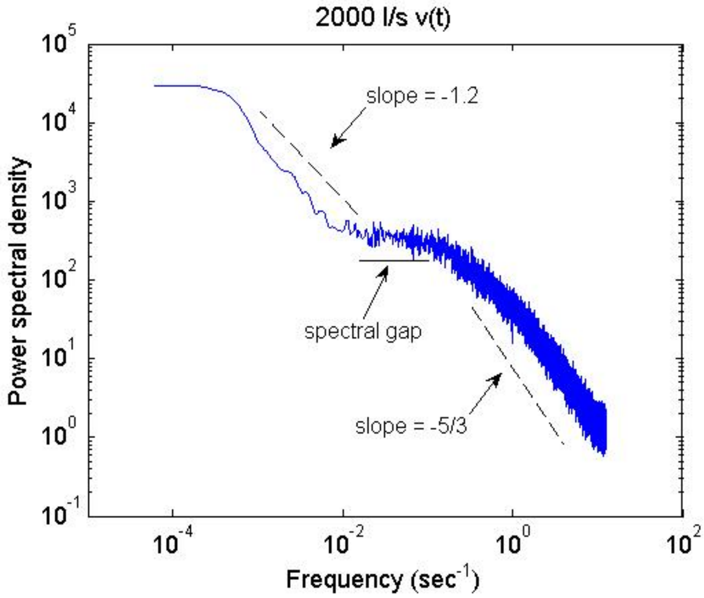

Figure 3. Power spectral density of velocity fluctuations for a discharge of 2000 l/s. The scaling at small scales is due to turbulence and at larger scales is affected by bed topography.

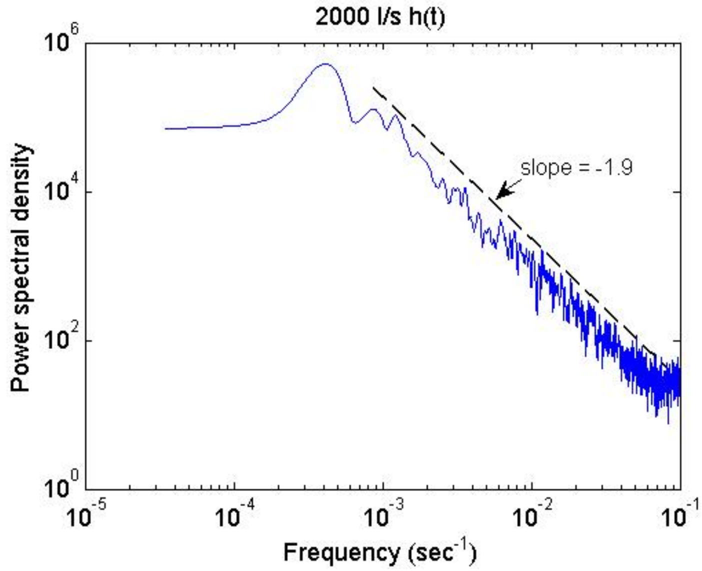

Figure 4. Power spectral density of bed elevation for a discharge of 2000 l/s.

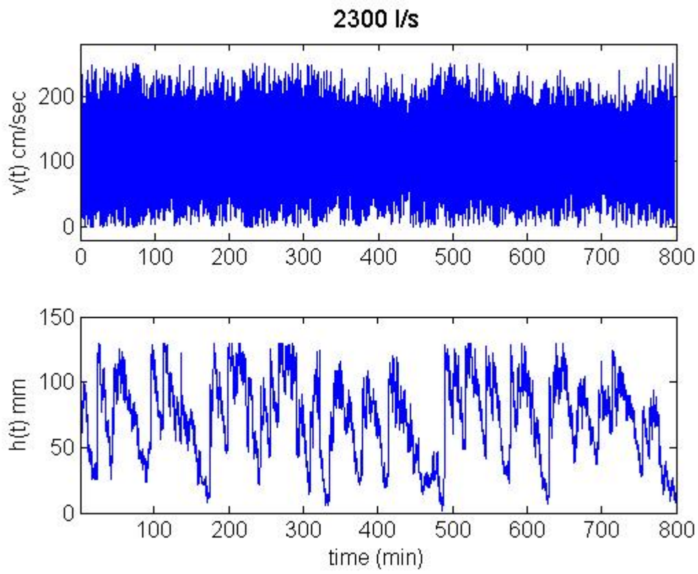

Figure 5. Velocity fluctuations measured at the downstream end of the channel at a frequency of 200 Hz (top) and the corresponding bed elevation measured at a frequency of 0.2 Hz (bottom) for a flow discharge of 2300 l/s.

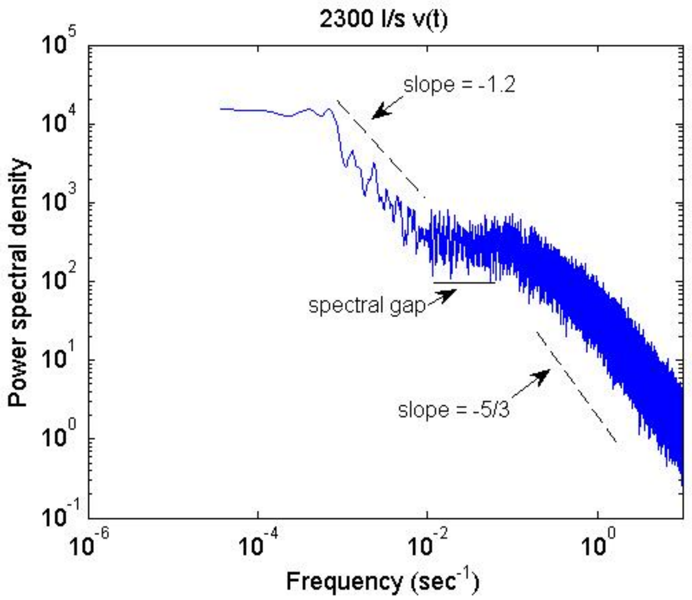

Figure 6. Power spectral density of velocity fluctuation for a discharge of 2300 l/s. The scaling in low scale is due to turbulence and in high scales is affected by bed topography. (Note that in this case the velocity fluctuations were measured at a frequency of 200 Hz).

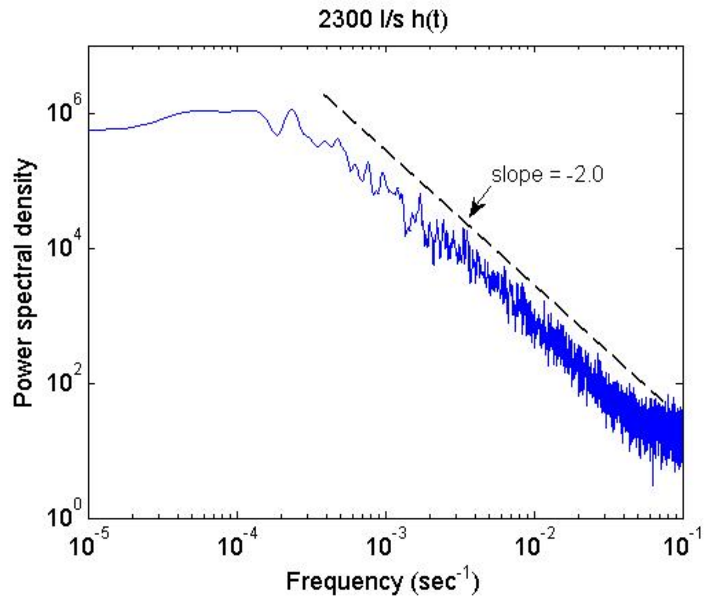

Figure 7. Power spectral density of bed elevation for a discharge of 2300 l/s.