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Hydraulic geometry of cohesive channels undergoing base level drop

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ABSTRACT

This study extends earlier contributions on dynamic adjustments of fluvial channels to base level changes. We have investigated an *in situ* response of self-formed cohesive channels to a base level drop, conditions analogous to a gradual change in uplift and/or climate. Empirical hydraulic geometry equations for clayey-cohesive natural streams are presented using data from eight channels draining perennial brackish springs and discharge into the Dead Sea. Investigation of downstream variations in gradient and stream power relations suggests existence of three distinct reaches in which channel adjustment to base level drop is shared inequitably among hydraulic geometry variables. Values of the flow velocity exponent m are low ($0.11 \leq m \leq 0.24$), the mid-channel reach having the lowest exponent. The depth exponent f has the lowest value ($f \approx 0.3$) for the uppermost channel reaches, the rest having higher values ($f \approx 0.4$). The smallest width exponent ($b = 0.35$) characterizes the upper reaches. These values and their spatial distribution exhibit a regular pattern. We show that the lowermost channel reach adjusts by profile steepening and channel narrowing ($f > b$); the prevailing mechanism in the mid-channel reaches is lateral (width) adjustment, cross sections transiently transforming toward equilibrium; the uppermost reaches have wide and shallow channel cross sections because of series of bank collapses and resultant sediment aggradation, bringing rise to decreased local gradient, forcing further channel widening. The results of this study not only allow inference about how cohesive channels regulate their geometry, but also reveal the means by which hydraulic forces overcome substrate resistance, adjusting slope and channel dimensions and, as such, have implication for reach-scale channel morphology and models of stream power.

1. Introduction

The concept of hydraulic geometry, as introduced by Leopold and Maddock (1953) relates water-surface width, w, average depth, d, and average velocity, v, to discharge, Q, at a given stream cross section or reach in the form of power-law equations with locally-determined empirical coefficients a, c, k, and exponents b, f, m:

$$w = aQ^b \tag{1.1}$$

$$d = cQ^f \tag{1.2}$$

$$v = kQ^m \tag{1.3}$$

where exponents b, f, and m represent the rate of change of the hydraulic variables w, d, and v, respectively, with Q; coefficients a, c, and k are scaling factors that define the values of w, d, and v when $Q = 1$.

The underlying assumption of the relationships given by power law relations (described by Eqs. (1.1)–(1.3)) is that there is a certain discharge or dominant flow controlling channel dimensions (Knighton, 1987; Rhoads, 1991). Therefore, a channel adjusts its form in response to changes in discharge. The power-law form of Eqs. (1.1)–(1.3) essentially dictates that

$$a + c + k = 1 \tag{2}$$

and that

$$bfm = 1 \tag{3}$$

Consequently, Eqs. (1.1)–(3) suggest that rivers tend to develop in such a manner that generates an approximate equilibrium between channel dimensions and the water discharge they transport.

Discharge may vary either in time at a given channel cross section (at-a-station or at-a-point hydraulic geometry), or in space along a given river channel (downstream hydraulic geometry). Although at-a-station hydraulic geometry and downstream hydraulic geometry are represented graphically using similar methods, they are essentially different in terms of both underlying mechanics and applications (Ferguson, 1986; Clifford, 1996). At-a-station hydraulic geometry describes hydraulic geometry of a given section as a function of return period of flow; downstream hydraulic geometry involves spatial variation in channel form and process at a constant frequency of flow.

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Consequently, because at-a-station relations derive from a variety of channel geometries, they also exhibit variability in their hydraulic geometry relations (Park, 1977). In contrast, some degree of consistency of hydraulic geometry has been demonstrated when considered in the downstream sense. The relations (Eqs. (1.1)–(3)) applied to field- and laboratory- based observations for a variety of environments are strikingly close to the average values reported in the literature for channels in the equilibrium state, where b ~ 0.5, f ~ 0.4, and m ~ 0.1 (Leopold and Maddock, 1953; Singh et al., 2003). The majority of hydraulic geometry relations were obtained for alluvial (e.g., Andrews, 1984; Cao and Knight, 1998; Lee and Julien, 2006) and bedrock (Montgomery and Gran, 2001; Turowski et al., 2008; Wohl and David, 2008) rivers. Few studies have examined cohesive-bed rivers. Helmio (2004) assessed at-a-station hydraulic geometry for lowland depositional cohesive-bed rivers, and Ebisa Fola and Rennie (2010) presented downstream hydraulic geometry relations for clay-dominated cohesive bed rivers. These studies suggest that cohesive channels have distinguishable hydraulic geometry exponents that deviate considerably from the mean values (i.e., those for alluvial channels) and emphasize the importance of the substrate into which the channels incise in determining fluvial morphology.

In studies of river adjustments, evolution of topography as a result of channel response to tectonic forcing has become a central concept in geomorphology (e.g., Snyder et al., 2000, 2003). However, similar to the studies of hydraulic geometry, the majority of recently developed models of channel evolution characterize commonly observed styles and sequences of alluvial (Leopold and Maddock, 1953; Knighton, 1998; Wohl, 2000; Whipple, 2004) and bedrock (Seidl and Dietrich, 1992; Montgomery and Gran, 2001; Finnegan et al., 2005) channels.

Rivers undergoing base level change have been reported to respond by degradation, aggradation, changes in channel pattern and geometry, or a combination thereof (e.g., Begin et. al., 1981). Lowering the base level was also shown to cause initiation of gullies (e.g., Burkard and Kostaschuk, 1995), the development of channel terraces (e.g., Hassan and Klein, 2002), knickpoint migration (e.g., Parker and Izumi, 2000), and acceleration of bank erosion (e.g., Robbins and Simon, 1983). However, no field-based study has documented sequential, temporal adjustments of hydraulic geometry parameters in cohesive channels in response to uplift/base level drop, a transient change that is `hidden` by the power law. This is mainly attributed to the fact that the study of the erosion of cohesive beds is not as advanced as that of noncohesive beds. Yet, it has been established that the erodibility of noncohesive sediments mainly depends on physical properties of particles such as the size, shape, density, porosity, and fall velocity (e.g., Yallop et al., 1994; Rehmann and Soupir, 2009), whereas the resistance to erosion of cohesive sediment depends heavily on the strength of the cohesive bond binding the particles (e.g., Mehta, 1986; Mitchener and Torfs, 1996; Mazurek et al., 2001; Zhu et al., 2008). Unlike coarse-grained sediments, cohesive sediments are not classified by grain size or grain size distribution. The primary flow-induced parameter characterising the applied erosive energy (Hanson and Simon, 2001) and a potential hydraulic-detachment mechanism operating longitudinally along the streambed (Simon et al., 2001) is bed shear stress (τ_b). To estimate downstream changes in τ_b within steady, uniform flow conditions, τ_b is usually expressed as a function of water discharge Q, channel width W and gradient S (e.g., Snyder and Kammer, 2008):

$$\tau_b \propto Q^{3/5} W^{-3/5} S^{7/10} \tag{4}$$

However, this relationship ignores the prominent effect of varying channel geometry by assuming a constant width-to-depth ratio.

The current study aims to document sequential changes in the scaling behaviour of hydraulic geometry in natural cohesive channels following their adjustment to base level drop. The decrease in water level in the Dead Sea has caused rapid exposure of marshy-clayey lake beds, followed by relatively rapid channel incision. These channels provide an *in situ* laboratory to investigate the transient response of natural channels, which is vital for testing and validating models. We capture the *in situ* response of fluvial systems to these disequilibrium conditions for a wide range of discharges. Our investigation takes two steps beyond previous studies by (i) observing the *in situ* initiation and development of self-formed channels in cohesive sediments due to base level drop, hence allowing the channel to co-evolve and adjust autonomously; and (ii) measuring the changes in cross-sectional geometry and *in situ*-derived spatial distribution of hydraulic geometry parameters, hence eliminating the need for *a priori* constraints on channel geometry.

2. Study area

This study was performed in the northern part of the western shore of the Dead Sea (Fig. 1A). The Dead Sea and its Late Pleistocene precursor, Lake Lisan, experienced large fluctuations during the Late Quaternary in response to climatic variations (Bowman, 1971; Karniel and Enzel, 2006). The Dead Sea level drop from −392 m in 1945 to −426 m in 2012 (Nof et al., 2012) was intensified during the 1980s and is since maintained at a rate of ca 1 m y^{-1} (Bookman et al., 2006).

The effect of this abrupt retreat on the adjustment and reorganization of drainage systems was investigated for bedrock and alluvial channels (e.g., Begin et al., 1981; Haviv et al., 2006; Ben Moshe et al., 2008; Bowman et al., 2010; Storz-Peretz et al., 2011). In this study we focus on the area (ca 4 km^2) where perennial brackish springs derive from the Cretaceous mountain aquifer and discharge into the sea (Fig. 1B). Ein-fesh'ha springs flow eastward along a wide, up to 1-km strip located west of the present shore. These springs owe their location and size not merely to the nearness to the main fault but to synclinal (Laronne-Ben Itzhak and Gvirtzman, 2005) and smaller fault diversions of groundwater (Mallast et al., 2011). The overall discharge of the spring-fed channels along this area was estimated to be 8×10^7 m^3 $year^{-1}$ (Rofe, 2003). Rapid Dead Sea level lowering has affected the state and position of the springs, initiating massive exposure of marshy clayey units in the northern part of the Dead Sea. Consequently, the initiation of the spring-fed, self-formed cohesive channels occurs by incision into newly exposed cohesive lacustrine sediments following the drop of the Dead Sea level. Continuously forming new channel reaches therefore undergo erosion, whereas the slope of the exposed strata (the former underwater topography - bathymetry) remains very low and almost constant (0.023–0.025).

The channels (bed and banks) consist of laminated clay to silt-sized, clastic deposits comprised of 15% sand, 72% silt, and 13% clay (Shapira, 2006). The mineralogical assemblage of clays is represented by 30–50% kaolinite, 30-50% illite-smectite, and 5–10% (each) of illite and paligorskite (Nathan et al., 1990). Geochemical analyses show that the deposits contain a relatively low (between 0.6 and 0.8%) organic matter content, which is mostly derived from the land masses surrounding the lake (Nissenbaum et al., 2002).

Use of standard soil mechanics definitions and testing approaches for the deposits have been found inapplicable (Frydman et al., 2008), particularly in view of their exceptionally high salt content and since the ground water table in the Dead Sea area is commonly close to the surface such that the deposits are saturated.

Liquid limit values are very low, in part because the material does not dry sufficiently and because salts affecting the surface activity of the particles (Frydman et al., 2008). Plastic limit values suffer from the same problems. Triaxial tests performed on undisturbed samples of the material showed no systematic relation between strength parameters of the material and dry unit weight. However, one strength envelope yielded very low cohesion and an effective friction angle of 34°, which was interpreted as a discrepancy of total stress during testing (Frydman et al., 2008).