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The morphodynamic equilibrium state of a river in backwater dominated reaches

Liselot Arkesteijn[1*], Robert Jan Labeur[1], Astrid Blom[1]
[1] *Delft University of Technology, Faculty of Civil Engineering and Geosciences, Department of Hydraulic Engineering, P.O. Box 5048, 2600 GA, Delft, the Netherlands*
** Corresponding author; e-mail: e.c.m.m.arkesteijn@tudelft.nl*

Introduction

When rivers are forced by statistically invariant boundary conditions (i.e. an upstream water discharge, upstream sediment discharge and downstream base level that fluctuate around constant mean values), and are not subject to any forcing with a temporal trend (e.g. no uplift/subsidence, no sea-level rise), they tend to a morphodynamic equilibrium state over time. Due to continuously changing boundary conditions a river may never reach its mean equilibrium state, yet it will tend to it continuously, and if the boundary conditions change at a sufficiently slow pace, the river may be in a quasi-equilibrium state. Therefore, studying the equilibrium state of a river may help us to better understand the long-term trends that are observed in natural rivers, such as for instance the ongoing bed degradation in the Dutch Rhine.

Available models used to predict the morphodynamic equilibrium state are mainly analytical ones that start from the assumption that there is always normal flow, during all stages of an imposed upstream hydrograph (Prins, 1969; Blom et al., in preparation). This means the hydrograph may include variable flow rates due to for instance flood waves, yet the hydrodynamic state of the river is modelled as a sequence of consecutive normal flow regimes. Variable flow rates, tidal forcing and spatial variations in, for instance, river width, however, can induce backwater effects, also in the equilibrium state. Here we propose an efficient model that describes the river's behaviour also outside of the normal flow zone, in the so-called backwater segment (e.g. Nittrouer et al. 2012). The efficiency of this model results from the approach to solve for the equilibrium in a space-marching solution procedure (i.e. a backwater alike solution procedure), rather than using a time-marching model where long simulation times (e.g. 1000 years) are required before an equilibrium situation is reached.

Definition of equilibrium

In the equilibrium state, the expected rate of change over time of the bed level and bed texture is zero. As a direct consequence, the expected or mean sediment load (per grain size fraction) at each cross section is the same as the average sediment load (per fraction) supplied from upstream (under the assumption that abrasion can be neglected). This means that the bed level, texture and actual sediment transport rates can still vary over time, as long as the fluctuations average out over a sufficiently long period (De Vries, 1993). Here 'sufficiently long' refers to the period for which the boundary conditions are statistically invariant, e.g. a few years in which the observed discharges describe the full probability distribution of water discharges reasonably well. This notion of a stochastic equilibrium is illustrated in Fig.1.

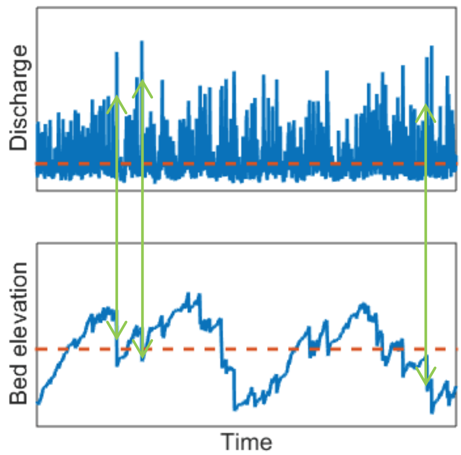

Figure 1: Stochastic equilibrium of the river for a conceptual case. Over time the river bed level fluctuates around a constant mean value, where the changes in bed level are correlated to the variable flow rates. At this specific location, large discharge peaks result in sudden bed erosion as indicated by the green arrows.

Model description

A local approximation

For rivers with a subcritical flow regime that are debouching in a large lake, sea or ocean, the base level can be considered constant in time, or varying with the tides. We may therefore suppose that at the downstream end, the statistics of the water discharge, and the water surface elevation (e.g. constant downstream base level) are imposed. The equilibrium requirement and the additional assumption that the temporally varying morphodynamic state (in equilibrium) can be approximated by the mean equilibrium state, then allow us to compute a local equilibrium state. This means we can compute the mean bed level, mean bed surface texture and derive all local flow variables, such as the flow depth, flow velocity, and Froude number during the various stages of the hydrograph, that on the long term facilitate the transport of exactly the average sediment load per grain size fraction.

Approximation of a single branch

Under the assumption that the variable flow can be treated as a sequence of steady flows (i.e. the backwater approximation), the full longitudinal equilibrium profile can be computed. We start downstream at the cross section where the equilibrium is known and compute solutions at the other cross sections by marching in upstream direction. The water surface slopes in the downstream cross section during the various stages of the hydrograph can be expressed as a function of the local friction slopes, the Froude numbers, and the mean bed slope. We note that only the mean bed slope requires information from a cross-section upstream (i.e. the bed level upstream), while the others can be estimated from the information that is available at our known (downstream) cross section. However, when we impose as equilibrium requirement that the spatial gradient of the expected sediment load (per grain size fraction) is zero, we introduce an extra set of equations which can be manipulated in such a way that they provide expressions for the approximation of the bed slope and the rate of change of bed surface texture, as a function of the local flow variables and the local domain characteristics (such as river width, friction, bed texture and porosity). This leads to a system of first order differential equations that describes the rate of change of all local hydrodynamic and morphodynamic variables in space. The equilibrium state can then be found by numerically integrating this system of equations along the river long-profile in upstream direction, using for instance an Euler forward method. Sufficiently far upstream of the backwater effects, the solution of our model reduces to existing analytical equilibrium models that provide expressions for the mean bed slope and mean surface texture under the assumption of normal flow.

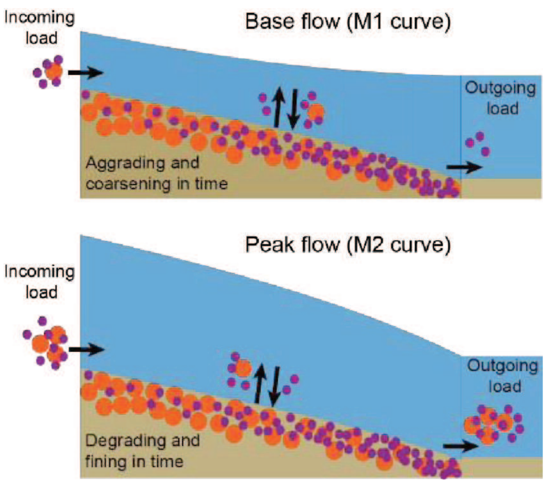

Figure 2: Convex bed profile and downstream fining in the equilibrium state.

Approximation of a river system

Local water or sediment extractions, variations in river width, confluences and tributaries can cause a sudden change in water and/or the required mean bed slope that is required to transport the average sediment load. Under the assumption that the water level is continuous, we can compute the mean bed level at the upstream side of the perturbation that satisfies the equilibrium requirement at the upstream reach. Please note that this can lead to a discontinuity in the bed profile, and that for tributaries and confluences there are multiple upstream branches. Once the upstream bed level(s) are known, the solution procedure can be continued (per branch). Dealing with bifurcations is less trivial and at this moment not included in the model yet.

Time reconstruction

When the mean equilibrium state and the hydrodynamic steady state during each discharge are known, we can compute the gradients in sediment transport during each discharge stage. After that, by ordering the erosion/deposition amounts per discharge in the order of occurrence of discharges, the bed level and bed texture fluctuations can be mimicked. This leads to an approximation of the bed level and bed texture change in time.

Effect of variable flow

The effect of the variable flow does not only introduce dynamic behaviour, it also leads to a different mean equilibrium state of the system in comparison to the mean equilibrium state under normal flow conditions. Fig. 2. illustrates the equilibrium state of a river section with constant width and a constant downstream base level. The alternating backwater effects lead to a mean convex upward profile, and in most cases a moderate fining of the bed surface texture in downstream direction.

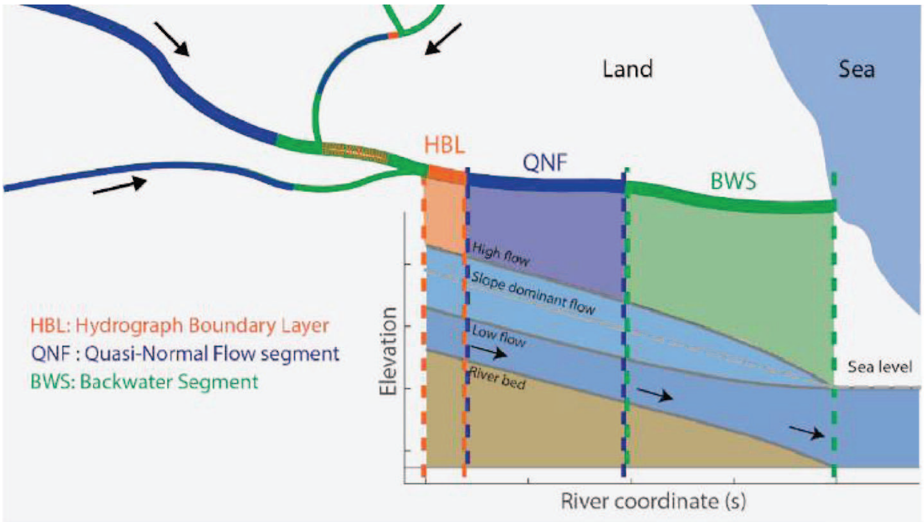

Figure 3: A river system in stochastic equilibrium subdivided into sections, classified as hydrograph boundary layer, quasi-normal flow segment, backwater segment, or a mixture of those.

Validation and discussion

In addition to the upward propagating perturbations caused by backwater effects, the local changes in river parameters also induce perturbations that are propagating in downstream direction in the 'hydrograph boundary layer' (e.g. Parker et al. 2004). These oscillations dampen out in streamwise direction, and the river adjusts toward a state where normal flow is prevailing during all stages of the hydrograph, while the sediment load has adjusted to the 'normal flow load distribution' (Blom et al., in preparation). Here the bed level does not change in time with the varying flow. A river system can therefore be considered to consist of zones where the behaviour is either best characterised as dominated by 1) downward propagating perturbations in bed level or bed texture (hydrograph boundary layer), 2) the absence of significant temporal variations in bed level and bed texture (quasi-normal flow segment), or 3) dynamic behaviour induced by backwater effects (backwater segment). Up- and downstream of each perturbation along the river section (e.g. varying width, or a tributary), a backwater segment and hydrograph boundary layer occur, which may overlap when the distance between two subsequent perturbations is too short for the oscillations to dampen out. This is illustrated in Fig. 3.

In our model, we do not incorporate the behaviour in the hydrograph boundary layer, as a solution procedure in upstream direction implies that downstream propagating information cannot be included. Also, since we formulated our model under the backwater-approximation, the morphodynamic effect of the dampening and hysteresis effect of a flood wave are not included.

The model has been validated for simple reaches where the zones do not overlap, using a numerical model that discretizes the Saint-Venant-Hirano model and that is able to predict the full dynamic behaviour as a reference solution. For a wide range of parameter settings, our proposed model is able to capture the behaviour in the quasi-normal flow zone and backwater segment very well. Furthermore the reduction in computation time is significant. While our space-marching model requires only a few minutes, the Saint-Venant model requires a few days, dependent on the quality of the initial condition. In addition, for the latter model it is cumbersome to define whether an equilibrium state is reached. Future work aims at extending the validation to cases where the hydrograph boundary layer and backwater segment do overlap.

Acknowledgements

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