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Global suspended sediment and water discharge dynamics between 1960 and 2010: Continental trends and intra-basin sensitivity

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ABSTRACT

Establishing a quantitative description of global riverine fluxes is one of the main goals of contemporary hydrology and geomorphology. Here we study changes in global riverine water discharge and suspended sediment flux over a 50-year period, 1960–2010, applying a new version of the WBMsed (WBMsed v.2.0) global hydrological water balance model. A new floodplain component is introduced to better represent water and sediment dynamics during periods of overbank discharge. Validated against data from 16 globally distributed stations, WBMsed v.2.0 simulation results show considerable improvement over the original model. Normalized departure from an annual mean is used to quantify spatial and temporal dynamics in both water discharge and sediment flux. Considerable intra-basin variability in both water and sediment discharge is observed for the first time in different regions of the world. Continental-scale analysis shows considerable variability in water and sediment discharge fluctuations both in time and between continents. A correlation analysis between predicted continental suspended sediment and water discharge shows strong correspondence in Australia and Africa (R^2 of 0.93 and 0.87 respectively), moderate correlation in North and South America (R^2 of 0.64 and 0.73 respectively) and weak correlation in Asia and Europe (R^2 of 0.35 and 0.24 respectively). We propose that yearly changes in intra-basin precipitation dynamics explain most of these differences in continental water discharge and suspended sediment correlation. The mechanism proposed and demonstrated here (for the Ganges, Danube and Amazon Rivers) is that regions with high relief and soft lithology will amplify the effect of higher than average precipitation by producing an increase in sediment yield that greatly exceeds increase in water discharge.

1. Introduction

Quantifying riverine sediment flux and water discharge is an important scientific undertaking for many reasons. Water discharge is a key component in the global water cycle affecting our planet's climate (Harding et al., 2011), ecology (Doll et al., 2009) and anthropogenic activities (e.g. agriculture, drinking water, recreation; Biemans et al., 2011). Quantifying sediment flux dynamics is a fundamental goal of earth-system science for its role in our planet's geology (Pelletier, 2012), biogeochemistry (Vörösmarty et al., 1997; Syvitski and Milliman, 2007) and anthropogenic activities (Kettner et al., 2010). Our quantitative understanding and predictive capabilities of global river fluxes are lacking (Harding et al., 2011). This is, in part, due to the multi-scale nature of the processes involved (Pelletier, 2012) and the inadequacy in global gauging of rivers (Fekete and Vörösmarty, 2007). Availability of measured river fluxes is decreasing globally (Brakenridge et al., 2012) particularly for sediment (Syvitski et al., 2005). Sediment fluxes to the oceans are measured for less than 10% of the Earth's rivers (Syvitski et al., 2005) and intra-basin measurements are even scarcer (Kettner et al., 2010).

Numerical models can fill the gap in sediment measurements (e.g. Syvitski et al., 2005; Wilkinson et al., 2009) and offer predictive or analytical capabilities of future and past trends enabling the investigations of terrestrial response to environmental and human changes (e.g. climate change; Kettner and Syvitski, 2009). Despite advances made in recent years (e.g. Kettner and Syvitski, 2008; Pelletier, 2012; Cohen et al., 2013) simulating global riverine fluxes remains challenging.

Climate change during the 21st century is projected to alter the spatio-temporal dynamics of precipitation and temperature (Held and Soden, 2006; Bates et al., 2008) resulting in natural and anthropogenically induced changes in land-use and water availability (Bates et al., 2008). Estimating the effect of these spatially and temporally dynamic processes warrants sophisticated distributed numerical models. Using past trends is perhaps the best strategy for developing these models and improving our understanding of the dynamics and causality within these complex systems.

Herein we present and validate an improved version of the WBMsed global riverine sediment flux model (Cohen et al., 2013). Cohen et al. (2013) showed that WBMsed can capture long-term average and inter-annual suspended sediment fluxes but tends to overestimate daily fluxes (by orders of magnitudes) during high discharge events and underestimate these during low flow periods. We found that these sediment flux miss-predictions are directly linked to miss-

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predictions of riverine water discharge, as the model's water routing approach did not limit the water transfer capacity of rivers. In other words, the model did not consider overbank flow and water storage in floodplains. For a natural river system, flooding not only limits the amount of water that can be transported over a certain period of time by a river but also provides a temporary reservoir that will resupply water back to the river days after the flood. The absence of such mechanism will result in a modeled river system that is overly responsive to runoff (i.e. overestimation during peak flow and underestimation during low flows) (Coe et al., 2008; Paiva et al., 2011; Yamazaki et al., 2011). Here we employ a floodplain reservoir component to store overbank flow at a pixel scale resulting in more realistic riverine flux predictions during peak and low flow conditions.

The new model is used to simulate water discharge and suspended sediment flux (at 6 arc-minute resolution) between 1960 and 2010. The results are used to analyze the yearly trends (normalized departure from mean) at both pixel scale and continental average. In this paper we focus our analysis on continental-scale interplay between suspended sediment flux and water discharge. A more focused analysis in three large basins (Ganges, Danube and Amazon) is preformed to explain discrepancies between water and sediment discharge, demonstrating an intriguing spatial–temporal interplay between lithology, topography and precipitation.

2. Methodology

2.1. The WBMsed v.2.0 model

WBMsed is a fully distributed global suspended sediment flux model (Cohen et al., 2013). It is an extension of the WBMplus global hydrology model (Wisser et al., 2010), part of the FrAMES biogeochemical modeling framework (Wollheim et al., 2008).

2.1.1. Water discharge module

The WBMplus model includes the water balance/transport model first introduced by Vörösmarty et al. (1989) and subsequently modified by Wisser et al. (2010). At its core the surface water balance of non-irrigated areas is a simple soil moisture budget expressed as:

$$dW_s/dt = \begin{cases} -g(W_s)\left(E_p - P_a\right) & P_a \le E_p \\ P_a - E_p & E_p < P_a \le D_{ws} \\ D_{ws} - E_p & D_{ws} < P_a \end{cases} \tag{1}$$

driven by $g(W_s)$, a dimensionless soil function:

$$g(W_s) = \frac{1 - e^{\left(-\alpha \frac{W_s}{W_c}\right)}}{1 - e^{-\alpha}} \tag{2}$$

where W_s is soil moisture, E_p is potential evapotranspiration, P_a is precipitation (rainfall P_r combined with snowmelt M_s), and D_{ws} is soil moisture deficit. Soil moisture deficit is the difference between available water capacity (W_c) and soil moisture. Available water capacity is dependent on soil and vegetation characteristics of each grid-cell (specified by input layers). The dimensionless empirical constant α is set to 5.0 following Vörösmarty et al. (1989).

Flow routing from grid to grid cell follows a downstream grid-cell tree topology (that allows the conjunctions of grid cells upstream but does not include diversions to, for example, river channel bars or multiple distributary channel deltas). Implementation uses the Muskingum–Cunge equation, a semi-implicit finite difference scheme to provide the diffusive wave solution to the St. Venant equations (ignoring the two acceleration terms in the momentum equation). The Muskingum–Cunge method is not the full-implementation of the diffusive wave approximation of the St. Venant equation. The Muskingum–Cunge solution includes a local diffusive effect within a single grid-cell, however it does not represent the diffusive effect between upstream and downstream grid-cells. Thus, the backwater effect caused by the water level rise in the downstream grid-cell is not represented in the calculation of the upstream discharge.

The equation is expressed as a linear combination of the input flow from current and previous time steps ($Q_{in\ t-1}$, $Q_{in\ t}$) and the released water from the river segment (grid-cell) in the previous time step ($Q_{out\ t-1}$) to calculate the new grid-cell outflow ($Q_{out\ t}$):

$$Q_{out\,t} = c_1 Q_{in\,t} + c_2 Q_{in\,t-1} + c_3 Q_{out\,t-1}. \tag{3}$$

The Muskingum coefficients (c_1, c_2, c_3) are traditionally estimated experimentally from discharge records, but their relationships to channel properties are well established. Detailed descriptions are provided in Wisser et al. (2010).

The new floodplain reservoir module (Fig. 1) adjusts daily water discharge for each grid-cell based on its bankfull discharge. When predicted water discharge ($Q_{out\ t}$) exceeds bankfull discharge (Q_{bf}) the "excess" water ($Q_{out\ t} - Q_{bf}$) will be stored in a virtual infinite floodplain reservoir and the new streamflow will equal bankfull discharge ($Q_{out\ t} = Q_{bf}$) (Fig. 1a). It should be noted that riverine water discharge cannot realistically exceed bankfull discharge and so the described equations below are an algorithmic rather than a physically-based solution. Once predicted water discharge is below bankfull again, water held in the floodplain reservoir will be reinjected to the river grid-cell. The volume of water returning to the river in a given time-step is proportional to the river grid-cell deficit from bankfull ($Q_{bf} - Q_{out\ t}$), i.e. very low river flows will result in greater reinjection of floodplain water (Fig. 1b). The changes in water discharge can be formulated as:

$$Q_{out\_aj} = \begin{cases} Q_{bf} & Q_{out\_t} > Q_{bf} \\ Q_{out\_t} + \left(Q_{bf} - Q_{out\_t}\right) b & Q_{out\_t} < Q_{bf} \end{cases} \tag{4}$$

where Q_{out_aj} is the adjusted river water discharge (m^3/s) and b is a daily delay fraction of water flow from the floodplain to the river ($b = 1$ translates to no delay (open flow)). For simplicity we assume here that $b = 1$ however a more complex description of b can be employed.

Bankfull discharge at a river segment is estimated using an approach modified from the river morphology module in the CaMa-Flood model (Yamazaki et al., 2011)

$$Q_{bf} = HWV_{bf} \tag{5}$$

where H is bank height

$$H = \mathrm{Max}\left[0.5\overline{Q}^{0.3}, 1.0\right] \tag{6}$$

where $\overline{Q}$ is long term average discharge, W is channel width

$$W = \mathrm{Max}\left[15\overline{Q}^{0.5}, 10.0\right] \tag{7}$$

and V_{bf} is bankfull flow velocity

$$V_{bf} = n^{-1} S^{-1/2} H^{2/3} \tag{8}$$

where n is Manning's roughness coefficient (0.03) and S, slope (m/m), is assumed to be constant. Here we used a slope value of 0.001 as a midpoint between very large, low-gradient rivers (e.g. Mississippi and Amazon with a slope of about 2.0×10^{-5}; Nittrouer et al., 2008 and LeFavour and Alsdorf, 2005) and steep headwater rivers (with gradients greater than 0.1; Chiari et al., 2010). A spatially explicit riverine slope description will improve the accuracy of this algorithm and is currently under development.

Additional approaches for estimating bankfull discharge were extensively tested. We have found that the Pearson III flood frequency estimator (using a 5-year flood frequency parameter) resulted in fairly realistic results. However this purely statistical methodology proved to