Diffusion modeling of recessional flow on central Amazonian floodplains

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[1] We present a continuity-based approach for calculating flow delivered to a main channel from an adjacent floodplain and use the values in a linear diffusion model to generalize fluxes across a floodplain. Using interferometric synthetic aperture radar (SAR) measurements of floodplain water level changes and the continuity equation, we demonstrate that flow rates are not the same throughout Amazonian floodplains. Also, rates of floodplain storage change are found to be least in areas of greatest distance from a main channel which suggests a long residence time. Linear diffusion modeling of floodplain drainage represents the composite behavior of flow through channels, swamps and lakes and provides a simple method of defining storage changes. The key parameter necessary for diffusion modeling is floodplain conductivity which can be constrained by a simple description of floodplain topography and measurements of temporal changes in floodplain water levels. Citation: Alsdorf, D., T. Dunne, J. Melack, L. Smith, and L. Hess (2005), Diffusion modeling of recessional flow on central Amazonian floodplains, Geophys. Res. Lett., 32, L21405, doi:10.1029/2005GL024412.

1. Introduction

[2] Floodplain inundation and drainage are important components of the sources and routing of flood waves along lowland river valleys. For example, based on Muskingum modeling, Richey et al. [1989] estimated that the Amazon mainstem exchanges about 25% of its average annual flow with the floodplain. It is difficult to measure the elevation of floodwaters over an entire floodplain, and especially to measure changes in water levels that contribute to the flux between a main channel and its floodplain. Thus, to estimate mass balances during the passage of a flood wave, the usual model of floodplain inundation and drainage envisions the water surface as horizontal and equal to the changing level of water in the axial channel [e.g., Richey et al., 1989].

[3] Recent work has shown that the situation is more complex [Mertes, 1997] and indicates the value of characterizing the spatial and temporal change of water surface elevations during both the inundation and drainage phases of flood seasons. For example, Alsdorf et al. [2000] used interferometric SAR measurements of water level changes across the central Amazon floodplain (i.e., $\partial h/\partial t$ where $h$ is the floodplain water surface elevation and $t$ is time) to show that the field of $h$ values is not horizontal and $\partial h/\partial t$ values are not equivalent to the associated main channel. Alsdorf [2003] spatially integrated the $\partial h/\partial t$ values to suggest that floodplain storage change errors of 30% may occur when assuming a horizontal water surface – a large error when considering the vastness of the Amazon floodplain.

[4] In this paper, we use interferometric SAR observations in the context of the continuity equation to construct hydrologic flux balances for select reaches of the Amazon, Purus and Negro floodplains in the central Amazon Basin. Variations amongst the flux balances allow us to describe the recessional flows. The $\partial h/\partial t$ values, combined with the flux balances, are then used in a linear diffusion hydraulic model of floodplain storage.

2. Floodplain Geomorphology and Hydrology

[5] The geomorphology of the Amazon and Purus reaches is distinct from that of the Negro (Figure 1). The Amazon and Purus rivers are surrounded by broad, low relief expanses that are inundated at the peak of the flood wave (~10 m). We consider the floodplains of the Amazon and Purus reaches as those areas that are inundated as indicated in Figure 1 (floodplain area, $A_f$, Table 1). They are bounded by land which is not flooded, called terra firme. Lakes ranging widely in size and shape abound on the floodplain [Sippel et al., 1992] and channels of various widths, depths, and degrees of boundary definition convey water along convoluted paths. Drainage across this complex landscape can be further impeded by floating grasses, trees and organic debris.

[6] In contrast, the lower Negro River lacks a large floodplain coupled with a single main channel (Figure 1). Flow is directed through a 15 km wide archipelago consisting of 500 m to 2000 m wide channels and islands. Terra firme uplands border the archipelago with elevations exceeding 40 m resulting in comparatively less inundated area per unit channel length than is found in the Amazon or Purus reaches ($A_f/L_f$, Table 1, $L_f$ is reach length). Most of the inundated area along the Negro consists of ria lakes located between a river channel and adjacent terra firme [Sippel et al., 1992] and we refer to this area as the Negro floodplain. These lakes are well defined in the SAR imagery such that flow delivered to the Negro River across...
Figure 1. Overlay of two mosaics of JERS-1 L-band SAR images over the central Amazon Basin acquired during the low-water period of late 1995 and during peak stage in 1996 [Rosenqvist et al., 2002]. White marks annually inundated areas; dark blues are always flooded; green is indicative of non-flooded areas [Hess et al., 2003]. Diagonal box locates the interferometric SIR-C swaths of Alsdorf et al. [2000]. Gauge locations are marked with arrows, study reaches span upstream floodway inputs and downstream outflows are time. Infiltration is neglected in equation (1) whereas mouth).

Table 1. Floodplain Reach Areas, Lengths, and Water Fluxes

<table>
<thead>
<tr>
<th>Reach</th>
<th>A_r</th>
<th>L_r</th>
<th>A_r/L_r</th>
<th>ΔS_f/L_f</th>
<th>q/L_f</th>
<th>Q(X_L, t)/L_f</th>
<th>Q(0, t)/L_f</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amazon</td>
<td>11800</td>
<td>310</td>
<td>38</td>
<td>15</td>
<td>0.45</td>
<td>2.2</td>
<td>18</td>
</tr>
<tr>
<td>Purus</td>
<td>6000</td>
<td>190</td>
<td>32</td>
<td>8.9</td>
<td>0.36</td>
<td>2.2</td>
<td>12</td>
</tr>
<tr>
<td>Negro</td>
<td>2000</td>
<td>280</td>
<td>7.1</td>
<td>3.0</td>
<td>0.082</td>
<td>2.2</td>
<td>5.3</td>
</tr>
</tbody>
</table>

Notes: Areas determined from Hess et al. [2003] image classifications. Nomenclature follows text: A_r in km²; L_r in km; A_r/L_r in km²/km; others in m³/s/km.

Figure 3. Trends drawn through the ∂h/∂t values are regarded as functions that describe the water level change throughout a floodplain for the recessional flow. Thus each floodplain reach is treated as having a smoothly changing ∂h/∂t with distance (local variations are likely more complex). Alsdorf [2003] used these trends to estimate ΔS_f of the Amazon floodplain reach as 4600 m³/s (+400 m³/s, −900 m³/s). Here, we use the same method to estimate ΔS_f for the Purus and Negro floodplains (1700 +440, −420 m³/s and 850 +250, −200 m³/s, respectively, ΔS_f/L_f Table 1). These are the first measurements of Amazon ΔS_f.

[9] The term q equals P-E integrated over the reach area. At six gauging locations within ~100 km of the SIR-C swath, the daily average rainfall from July peak stage to October was 4.6 mm/day. Assuming a typical evaporation rate of 3.6 mm/day for the area [e.g., Costa and Foley, 1999], we set q at 1.0 mm/day or 140 m³/s, 69 m³/s, and 23 m³/s for the Amazon, Purus, and Negro study reaches, respectively (q/L_f Table 1).

[10] The inflow from terra firme surrounding a floodplain, Q(X_L, t), is estimated from rainfall-runoff data. For example, flow from the terra firme to the Amazon floodplain in the Itapeua to Manaus reach was estimated by Richey et al. [1989] and Dunne et al. [1998] to range from 1200 m³/s in July to 500 m³/s in October; we use the average of 690 m³/s for Q(X_L, t) in each study reach (Q(X_L, t)/L_f Table 1). We assume that inflows from terra firme surrounding the Purus and Negro floodplains are similar to these model derived values for the Amazon floodplain.

[11] The flow delivered to a main channel from an adjacent floodplain is calculated from equation (1). During mid-recessional flow in October 1994, we obtained Q(0, t) values of 5500 m³/s, 2200 m³/s, and 1500 m³/s for the Amazon, Purus, and Negro study reaches, respectively (Q(0, t)/L_f Table 1). In the previous paragraphs, the time periods used for deriving each equation (1) quantity differ, thus these Q(0, t) values are intended as approximations for average recessional flow conditions.

[12] The interferometric observations in Figure 3 are the ∂h/∂t term in the mass continuity equation [e.g., Dingman, 1984],

\[ \frac{q}{W_f} = \frac{\partial Q}{\partial x} = L_f \frac{\partial h}{\partial t} \]

where L_f is the dimension perpendicular to the direction of floodplain flow (i.e., generally parallel to main channel flow and assumed to be equivalent to the reach length), x is a location anywhere between the floodplain boundaries, and W_f is the total distance between boundaries (Figure 2).
Nevertheless, Puru’s floodplain clearly defined as those of the Amazon and Negro reaches. However, the residence times per unit width (unit gauge values of Table 2. This further implies that water change rates are greater than distal rates (based on our equation (2)) demonstrates that proximal floodplain storage increased proximal floodplain water surface across the floodplain, which are built into one-dimensional flow routing, are in question.

4. A Linear Diffusion Flow Model

[15] We seek a simple model with limited parameterization to describe the broad features of floodplain storage and drainage. The model should be capable of incorporation in continental-scale water cycle models yet capture the basics of complex floodplain flow. The flow paths across the Amazon and Puru’s floodplains pass a volume of water moving through a myriad of channels and lakes with associated flooded forests and floating plants. We propose to represent the collective behavior of the water in these flow paths with a linear diffusion equation. Although flow across the Negro floodplain involves more discrete, less extensive flow paths, we apply the diffusion model there for comparisons.

[16] Previous uses of diffusion based modeling of floodplain flow have relied on high resolution digital elevation data (1 to 100 m² grid cells) and non-linear flow equations [Makhanov et al., 1999; Lal, 1998] to predict discharge, water depths and inundation extent. The model results generally matched gauge derived discharge and remote sensing measurements of inundation extent over small floodplains [Bates and De Roo, 2000], but only Makhanov et al. [1999] attempted to match temporal variations in water heights across a floodplain. We use the following one-dimensional, simple relationship to describe Q:

\[ Q(x,t) = -KL \frac{\partial h}{\partial x} \]  

(3)

where K is floodplain conductivity, Lh is the cross-sectional area perpendicular to the floodplain flux, and Q(x, t) is the flow across a floodplain (i.e., the x direction in Figure 2).

Substituting equation (3) into the mass continuity equation (equation (2)) yields the following equation of flow:

\[ q = W_f L_f \frac{\partial^2 h}{\partial x^2} + K h \frac{\partial h}{\partial t} \]  

(4)

Equation (1) expresses the relationship between changes in floodplain storage and the boundary flow terms indicated in Figure 2 which are necessary for solving equation (4).

[17] We developed a one-dimensional finite difference model to solve equation (4). The boundary conditions include a main channel with its water height, h(0, t), decreasing with time (i.e., measured from a gauge, Table 2), and an upland margin with a boundary inflow term (Q(xL, t) in Figure 2 and Table 1). For the initial conditions at peak stage in early July we assume that the water surface across the floodplain, h(x, 0) in Figure 2, is

\[ h(x, 0) = 0 \]  

for x ≤ 0 and h(x, 0) = h(x) for x > 0, where h(x) is the water surface measured at the main channel.

\[ h(x, t) = h(x, 0) - Q(xL, t) \]  

(1)

Table 2. Daily Changes in Main Channel Water Levels, 1994a

<table>
<thead>
<tr>
<th></th>
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<tbody>
<tr>
<td>Negro</td>
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<td>–6</td>
<td>–9</td>
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</tr>
<tr>
<td>Negro</td>
<td>Manaus</td>
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<td>–12</td>
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<tr>
<td>Amazon</td>
<td>Itapeau</td>
<td>–5</td>
<td>–5</td>
<td>–7.0</td>
<td>–10.1</td>
</tr>
<tr>
<td>Amazon</td>
<td>Manacapuru</td>
<td>–7</td>
<td>–13</td>
<td>–11.1</td>
<td>–10.1</td>
</tr>
<tr>
<td>Puru’s</td>
<td>Beruri</td>
<td>–12</td>
<td>–14</td>
<td>–10.4</td>
<td>–10.2</td>
</tr>
</tbody>
</table>

Notes: Drops in cm. Oct. 10 minus Oct. 4 changes (one week) and Oct. 10 minus Aug. 20 changes (half of the 101 days since peak stage) are an average of the encompassed one-day decreases.
equivalent to the stage in the main channel. Subtracting the bottom topography from \( h \), yields an average depth of floodplain flow at every \( x \) and \( t \) node. Therefore, in this application of equation (4), \( h \) is a function of \( x \) and \( t \), while \( K \), \( q \), and \( L \) are held constant.

[18] The two unknown parameters in equation (4), \( K \) and the general elevation of the floodplain surface, can be found by simultaneously (a) fitting the storage change predicted from equations (4) and (1) to the interferometric value of \( \Delta S \), and (b) fitting the daily differences in water heights derived from equation (4) to the interferometrically observed drops in water level (Figure 3). We used the Manacapuru gauge data for the main-channel boundary condition to model the flow across the Amazon floodplain between Itapeua and Manacapuru. Setting the elevation of the floodplain to a horizontal plane results in a model prediction that matches the interferometric observations and corresponding storage change estimate with a conductivity of 12800 km/day.

[19] To model the Purús floodplain reach between Aruma and the confluence with the Amazon, we used the Beruri gauge data and found that a ramp-and-flat bottom topography was necessary (Figure 3). The use of a horizontal plane in the Purús model resulted in predicted \( \partial h/\partial t \) values that either (a) matched the main channel but were more negative than the interferometric observations on the floodplain or (b) matched the floodplain but were less negative than the main channel water level change. The Purús floodplain diffusion model produces an approximate fit to both the drops and storage change with predicted \( K \) value of 6500 km/day, but notable differences do occur near the Purús channel. Water surface changes measured near the channel are smaller than predicted probably because the one-dimensional flow model does not include water forced overbank further upstream by backwater effects of the Amazon on the Purús [Meade et al., 1991]. For comparisons with the two larger floodplains, we also applied the diffusion model to the smaller Negro floodplain. Here, a steeper ramp than the Purús was required to fit the observations and a conductivity of 8500 km/day was derived.

[20] In this application of the diffusion equation, floodplain conductivity encompasses the effect on flow from various floodplain characteristics such as inundated area (some fraction of the floodplain may not be inundated), the topography created by various geomorphic features, channels of various sizes that fret the floodplain, and vegetation. Local variations in depths are also incorporated by the tradeoff between \( K \) and our use of general floodplain topography, wherein increasing the bottom slope and decreasing \( K \) can be used to maintain a constant \( Q \).

5. Conclusions

[21] Using just two parameters, conductivity and a basic description of floodplain topography, the 1D, linear-diffusion model is a simple method of characterizing complex floodplain flow. This approach revises the theoretical model employed in continental-scale water cycle models where water level changes in the main channel are instantly propagated laterally across the entire floodplain to one where this elevation change propagates across the floodplain as a 1D diffusion wave. The model captures the fundamental behavior of recessional flow as measured by satellite interferometric SAR.

References


