

# When interflow also percolates: downslope travel distances and hillslope process zones

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## Introduction

In hillslopes with soils characterized by deep regoliths, such as Ultisols, Oxisols, and Alfisols, interflow occurs episodically over impeding layers near and parallel to the soil surface such as low-conductivity B horizons (e.g. Newman *et al.*, 1998; Buttle and McDonald, 2002; Du *et al.*, In Review), till layers (McGlynn *et al.*, 1999; Bishop *et al.*, 2004), hardpans (McDaniel *et al.*, 2008), C horizons (Detty and McGuire, 2010), and permeable bedrock (Tromp van Meerveld *et al.*, 2007). As perched saturation develops within and above these impeding but permeable horizons, flow moves laterally downslope, but the perched water also continues to percolate through the impeding horizon to the unsaturated soils and saprolite below. Perched water and solutes will eventually traverse the zone of perched saturation above the impeding horizon and then enter and percolate through the impeding horizon. In such flow situations, only lower hillslope segments with sufficient downslope travel distance will deliver water to the riparian zone within the time scale of a storm. Farther up the slope, lateral flow within the zone of perched saturation will act mainly to shift the point of percolation (location where a water packet leaves the downslope flow zone in the upper soil layer and enters the impeding layer) down the hillslope from the point of infiltration. In flatter parts of the hillslope or in areas with little contrast between the conductivities of the upper and impeding soil layers, lateral flow distances will be negligible.

From Darcy's law, we can estimate the downslope travel distance for quasi-steady-state conditions within the saturated layer assuming Boussinesq (slope parallel) flow above the restrictive layer and normal flow (perpendicular to the slope) through the restrictive layer to unsaturated soils below (Figure 1). Using a downslope flow vector, we must employ an orthogonal normal vector and avoid Harr's (1977) mistake of adding non-orthogonal and non-independent gradient vectors in calculating flow directions. We assume that each layer is isotropic, and the impeding layer is saturated or nearly so and drains freely to an unsaturated layer below (pressure head equals zero at the base of the impeding layer). We assume diffuse porosity within the impeding layer so percolation is not restricted to randomly spaced discontinuities. With these assumptions generally applicable to hillslopes with deep regoliths and a low conductivity layer paralleling the soil surface, we will calculate downslope travel distances for reasonable combinations of slopes, conductivities, and impeding layer thicknesses; evaluate these distances with respect to typical slope lengths; map the downslope travel distances on watersheds with soil and slope conditions matching these assumptions; and discuss how coupling Boussinesq downslope flow and percolation affect our understanding of the potential zonation of lateral flow processes.

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the saturated layer above the impeding zone,  $N$ , we have calculated travel distances for  $N=1$  m. Maximum values of downslope travel distances can be estimated by setting  $N$  equal to the normal thickness of the topsoil. The high end of reported ranges for average forested topsoil conductivities is about 2000 mm/h, and the lower range of average conductivities for B horizons is around 1 mm/h, so we used conductivity ratios ranging from 1 (no layering and no saturated interflow) to 2000 (very high contrast between the topsoil and the impeding layer). The minimum value of the normal gradient is 1 (a thin saturated lens and a very thick impeding horizon), and an extremely high value would be around 9 ( $N=2$  m and  $C_n=0.25$  m). We have used slopes ranging from 1 (0.57°) to 100% (45°), and  $\sin \Theta$  thus ranges from 0.0099 to 0.71. For these ranges of downslope and normal gradients, possible gradient ratios range from 0.001 (1% slope with a very thin impeding layer) to 0.7 (100% slope with a normal gradient of 1). Within the resulting field of possibilities, downslope travel distances are small relative to hillslope lengths except for combinations of very high conductivity and gradient ratios (Figure 2). At the hillslope that we are studying at the Savannah River Site in the Sandhills of South Carolina (Du *et al.*, In Review), the maximum downslope travel distance is about 29 m (maximum surface conductivity of 2000 mm/h, clay conductivity of 5 mm/h, surface slope of 11%, normal clay thickness of about 2 m, and a maximum saturated zone of 1-m normal thickness above the impeding layer), meaning that interflow on this slope does not deliver water to streams or valleys but simply shifts the percolation of infiltrated water, at most, 30-m downslope. Many

interflow studies have been conducted in steep terrain underlain by low permeability rock and therefore feature high conductivity and gradient ratios and, thus, long interflow travel distances. In such cases, interflow travel distances can exceed 1000 m (Figure 2).

### Implications and Considerations

In rolling, highly weathered terrain where water perches upon a porous B horizon, downslope travel distance estimation predicts that interflow delivers water to the valleys from only the lower reaches of the slope and suggests zonation of the dominant hydrologic processes from ridgetop to stream. In such terrain, we can envision four dominant process zones (Figure 3). On the relatively flat ridgetops and intermediate benches, interflow is an unimportant process, and vadose zone hydrology is largely vertical. Over most of the slope, interflow acts to redistribute percolation of water and solutes, shifting the point of infiltration by the travel distance. Of course, the movement of solutes is complicated by dispersion, mixing with old water, sorption, and desorption, so this calculation addresses only the net movement of solutes. Only rainfall that falls on the lower slopes within the downslope travel distance of the valley may be delivered to the valley as saturated interflow. In this lower region of the slope, interflow may deliver water directly to the stream if the channel is cut to the base of the slope, but it is more likely that interflow acts to subsidize valley hydrologic processes (e.g. McGlynn and McDonnell, 2003; Jencso *et al.*, 2010). Interflow from the lower portions of valley-adjacent hillslopes can be considered an extension of the variable source area. Groundwater mounding in the alluvial aquifer or toe slope (Ragan, 1968; Sklash and Farvolden, 1979, Abdul and Gillham 1989) has been attributed to the proximity of the water table to the ground surface and the relative wetness of the soils there, but interflow contribution from the lower slopes can also contribute to the formation of groundwater mounds (Figure 3). The degree to which interflow subsidizes groundwater mounding *versus* saturation excess flow depends on the valley topography, particularly the existence and thickness of a terrace deposit above the floodplain. This process zonation exists even when the saturated lens is continuous from the ridgetop to the stream as observed by Detty and McGuire (2010).

To illustrate the variation of downslope travel distances over first-order watersheds with soils and topography matching the assumptions herein, we estimated and mapped the downslope travel distances on three first-order watersheds in the Sandhills of South Carolina located within the Savannah River Site and

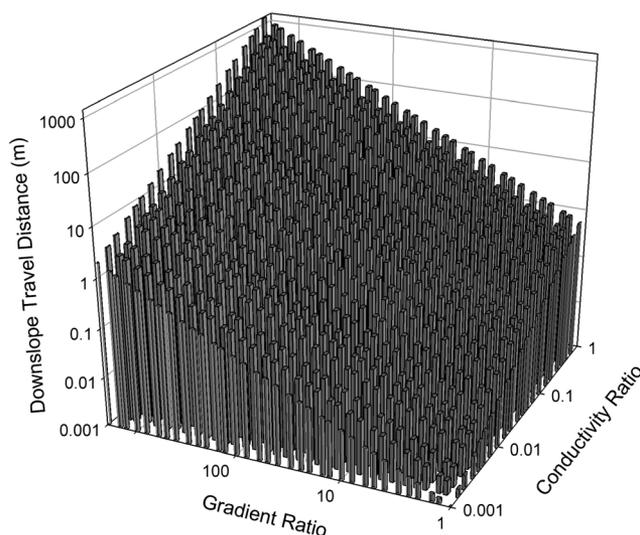


Figure 2. Downslope travel distances (metres) for all likely combinations of gradient and conductivity ratios. Calculated for a 1-m thick saturated layer above the impeding horizon ( $N=1$  m). Travel distances scale linearly with  $N$

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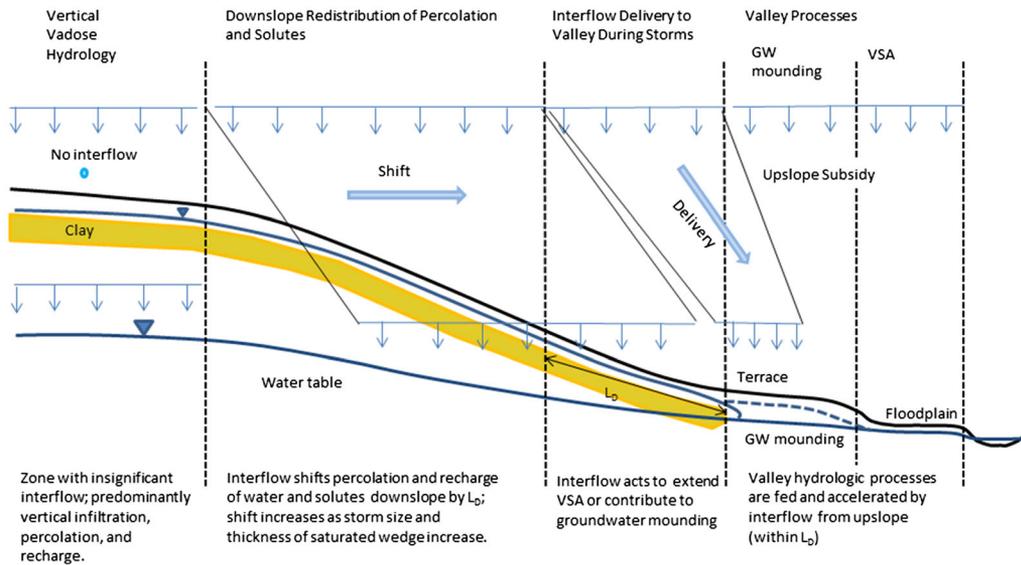


Figure 3. Implications of saturated interflow travel distance for the zonation of hillslope flow processes

described in Du *et al.* (In Review). Slopes were calculated based on the elevation difference and distance to the nearest stream cell generated from 10-m digital elevation model. Normal thickness of the impeding layer and conductivities was estimated from auger investigations and compact constant head permeameter measurements. Within these watersheds, maximum downslope travel distances are 36 m, and these higher values are generally located on steeper slopes directly adjacent to the stream valleys (Figure 4). Over most of the watersheds, downslope travel distances are quite short, less than 15 m. The travel distance maps also reveal the very low slopes within the riparian valleys

and wetlands where the valley hydrologic processes of saturation excess flow and groundwater mounding are dominant.

Travel distances scale directly with the thickness of the perched saturated zone above the impeding layer. While others have demonstrated the existence of rainfall/soil moisture thresholds for the initiation of interflow (e.g. Uchida *et al.*, 2005, Lehmann *et al.*, 2007), this analysis also shows that the upslope interflow contributing length to the stream valley increases with the thickness of the saturated layer, so the cumulative interflow contribution to the valley during a storm should increase nearly linearly with the size of the storm

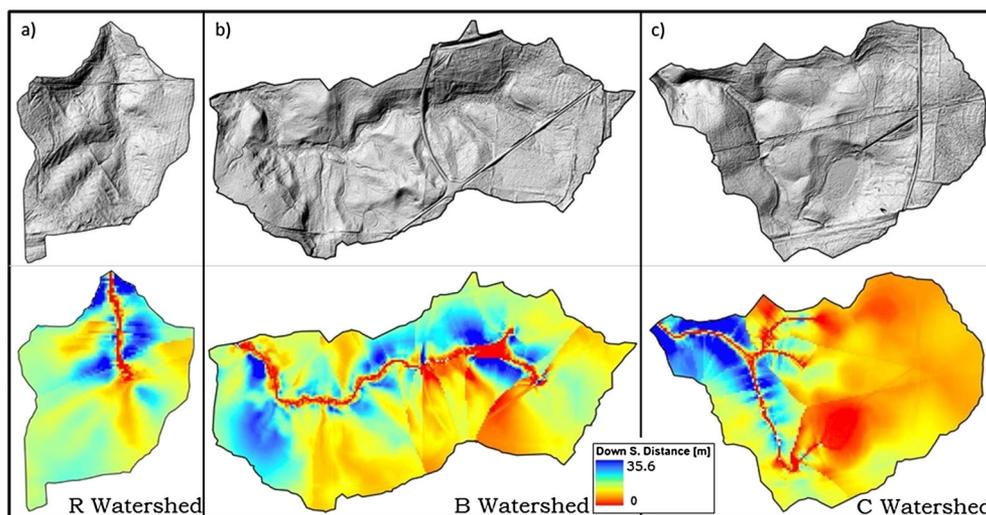


Figure 4. Map of downslope travel distances for three first-order watersheds in the Sandhills of South Carolina (Du *et al.*, In Review) based on average slope from each 10 × 10-m digital elevation model pixel and the nearest stream cell. A hillshade map based on light detection and ranging of each watershed is shown above the travel distance map. The C watershed features several Carolina Bay depositional wetlands in the upper part of the watershed, and these appear as circular or elliptical red areas

minus the threshold. While not a surprising result, this analysis clearly demonstrates that the relative importance of interflow on hillslopes falls on a continuum, leaving hydrologists with somewhat subjective decisions about when interflow must be considered and when it can be safely ignored. This analysis strongly suggests that many process-based models need to be modified to account for the longitudinal zonation of flow processes in hillslopes where such zonation is likely. The results also indicate that slope-parallel flow with negligible percolation is the dominant process in steep hillslopes with high conductivity contrasts between the topsoil and the impeding layer. In such cases, interflow models that ignore percolation (e.g. Troch *et al.*, 2003; Jackson and Cundy, 1992) are appropriate.

The limitations of our observational systems make it difficult to put interflow into context with other hillslope processes. Piezometer networks inform little about how far interflow moves. Because interflow often mixes with riparian water before entering streams, end-member mixing analysis cannot tell us how much of the 'old' riparian water arrived via interflow. Interception trenches measure only what is passing a given contour and impose flow-altering boundary effects. Tensiometers are temperamental and difficult to manage deeper in the soil profile. Given the variation of hillslope characteristics and the difficulties of monitoring interflow, generalizing the interpretation of results from different hillslope process studies is difficult. Consideration and mapping of downslope travel distances may be a tool for understanding and synthesizing results of interflow studies in different hillslope environments. Explicitly considering the normal flow component (percolation) as well as the downslope flow component allows the relatively easy identification and mapping of likely dominant process zones or hydrologic response units over entire watersheds (*sensu stricto* Devito *et al.*, 2005) based on soil characteristics and local topography. The resulting process zonation can be applied to water quality management issues and used to help define spatially variable riparian buffer widths as suggested by Dosskey *et al.* (2002 and 2013) and Rivenbark and Jackson (2004). However, as with most distributed modelling concepts, only the topography variables are easy to measure and map. Calculating downslope travel distances for perched interflow still requires knowledge of soil hydraulic conductivities and depths (e.g. Buttle *et al.*, 2004).

In summary, by considering percolation simultaneously occurring with saturated interflow, we can easily estimate the likely downslope travel distances of water and solutes before reaching the impeding layer and percolating through it. For many hillslope environments, particularly for rolling topography underlain by porous but impeding soil horizons, these downslope travel

distances are much shorter than slope lengths and suggest zonation of the importance and relevance of interflow along hillslope catenas. Interflow only contributes to and subsidizes valley hydrologic processes from the lower slopes, within the downslope travel distance from the riparian valley.

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