

1 **Evidence of subsurface control on the coevolution of**
2 **hillslope morphology and runoff generation**

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7 **Key Points:**

- 8 • We test theoretical predictions about the relationship between hillslope length and
9 relief, saturated area, and transmissivity
10 • Comparing two watersheds, we find lower transmissivity is associated with shorter
11 hillslopes and larger variably saturated areas
12 • Hydrogeomorphic modelling suggests that subsurface properties drive the coevo-
13 lution of differences between the sites

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Abstract

Topography is a key control on runoff generation, as topographic slope affects hydraulic gradients and curvature affects water flow paths. At the same time, runoff generation shapes topography through erosion, which affects landscape morphology over long timescales. Previous modeling efforts suggest that subsurface hydrological properties, relative to climate, are key mediators of this relationship. Specifically, when subsurface transmissivity and water storage capacity are low, (1) saturated areas and storm runoff should be larger and more variable, and (2) hillslopes shorter and with less relief, assuming other geomorphic factors are held constant. While these patterns appear in simulations, it remains uncertain whether subsurface properties can exert such a strong control on emergent properties in the field. We compared emergent hydrological function and topography in two watersheds that have very similar climatic and geologic history, but very different subsurface properties due to contrasting bedrock lithology. We found that hillslopes were systematically shorter and saturated areas more dynamic at the site with lower transmissivity. To confirm that these differences were due to subsurface hydrology rather than differences in geomorphic process rates, we estimated all parameters of a coupled groundwater-landscape evolution model without calibration. We showed that the difference in subsurface properties has a profound effect on topography and hydrological function that cannot be explained by differences in geomorphic process rates alone. The comparison to field data also exposed model limitations, which we discuss in the context of future efforts to understand the role of hydrology in the long-term evolution of Earth's critical zone.

Plain Language Summary

In many humid landscapes, runoff is generated by water that flows through the shallow subsurface from ridges to valleys, eventually emerging and draining to rivers. The greater the capacity of the subsurface to move water, the more water can collect before surface runoff begins. Surface water may cause erosion, which shapes ridges and valleys over millions of years. We previously developed a computer model based on these principles and showed that the subsurface capacity to store and transmit water affects both runoff generation and topographic evolution. Lower capacity results in more surface runoff and shorter, lower relief hillslopes, when all other factors are held constant. Here we tested this by comparing two watersheds that differ primarily in their bedrock composition, which affects subsurface water storage and transmissivity. We found that the low transmissivity site had more dynamic surface runoff and shorter hillslopes, supporting our predictions. We set up computer models for both sites, which suggested that subsurface differences are necessary to explain observed differences in runoff and topography. Finally, we discuss some key limitations of the model that could be improved upon in future attempts to understand how hydrology affects the long-term evolution of Earth's surface.

1 Introduction

1.1 Background

It has long been understood that there is a close, two-way connection between runoff and the topographic form of landscapes. Topography influences flow paths of water over the surface and through the subsurface and supplies the elevation component of hydraulic head, while erosion by water shapes landscapes over long timescales. Horton (1945) first suggested that there is something valuable to learn about how places work hydrologically by considering this coupling. While Horton's work focused on the role of infiltration excess overland flow in determining contributing areas and drainage network topology, Carlston (1963) suggested that we should also be able to learn something about groundwater-driven runoff based on channel spacing. However, the vastly different timescales of runoff and evolution of channel networks via erosion has made it challenging to study the co-

64 evolution of hydrological and geomorphic states and fluxes. As a result, hydrologists study-
65 ing runoff generation usually assume that landscape form is fixed, while geomorpholo-
66 gists studying landscape evolution usually assume hydrology can be reduced to a few pa-
67 rameters that capture how hydroclimate affects the efficiency of bedrock erosion and sed-
68 iment transport.

69 Recent advances in modeling and the availability of high performance computers
70 have allowed the coupling of hydrologic and geomorphic models that consider the evo-
71 lution of hydrologic and geomorphic states together. Litwin et al. (2022) used a shallow
72 aquifer model to generate saturation excess runoff from steady recharge, and used the
73 runoff to drive fluvial incision in a streampower-plus-diffusion landscape evolution model.
74 They showed that the thickness and permeability of the subsurface were important con-
75 trols on runoff, and as a consequence, the degree of drainage dissection and length of hill-
76 slopes. Litwin, Tucker, et al. (2023) extended this model to examine the emergence of
77 variable source area hydrology, adding stochastic precipitation and a simple represen-
78 tation of the vadose zone to the prior model to capture more realistic hydrologic dynam-
79 ics. Again, the thickness and permeability were key controls on both the morphology and
80 hydrological function of the coevolved landscapes. They showed that landscapes with
81 efficient subsurface drainage and large water storage capacity had less variable and smaller
82 saturated areas than those that had poor subsurface drainage, and therefore generated
83 less storm runoff. This difference in runoff response has implications for geomorphology
84 as well. Decreasing the spatial extent of runoff decreases the extent of fluvial erosion,
85 which decreases the degree of drainage dissection. Litwin, Tucker, et al. (2023) also found
86 an emergent relationship between runoff and morphology. Specifically, the fraction of quick-
87 flow relative to total discharge scaled inversely with the dimensionless hillslope relief in
88 the watershed. This relationship supported prior predictions (Dunne, 1978) that steeper
89 landscapes (with more transmissive soils) generated more runoff via subsurface flow, while
90 landscapes with gentle topography (and thinner less transmissive soils) generated more
91 runoff via saturation excess.

92 While these numerical results indicate that the subsurface is a key link between to-
93 pography and runoff generation, it is unclear whether these relationships appear outside
94 idealized models. While field studies have shown that subsurface properties and topog-
95 raphy have effects on hydrologic function (e.g., Prancevic & Kirchner, 2019; Jencso &
96 McGlynn, 2011), relationships between subsurface properties and topography remain elu-
97 sive (Luo et al., 2016; Sangireddy et al., 2016), let alone unambiguous evidence that the
98 link between them is the result of coevolution (Yoshida & Troch, 2016). This lack of clear
99 relationships is to be expected because hydrology, conditioned by climate, is only one
100 connection between the subsurface and topography. Other controls include lithology and
101 tectonic setting, which affect the styles and efficiencies of weathering and sediment trans-
102 port, and vegetation, which alters subsurface properties and sediment transport efficiency
103 through root growth, and hydrologic partitioning through evapotranspiration (Brantley
104 et al., 2017; Collins & Bras, 2010). If the subsurface connects topography and runoff gen-
105 eration despite all of this complexity, catchment coevolution may be a useful tool for un-
106 derstanding and predicting hydrological function (Troch et al., 2015).

107 1.2 Approach

108 If a signature of coevolution between topography and hydrological function exists,
109 we will be most likely to find it where we can isolate the hydrological effects from other
110 influences. We selected two sites where contrasting lithology results in a strong contrast
111 in subsurface properties, but climatic and tectonic histories are similar because of their
112 proximity. Our first site, Druids Run, is underlain by serpentine bedrock that forms thin
113 rocky soil, while the second site, Baisman Run, is underlain by schist that weathers to
114 form deep, permeable soil and saprolite. Assuming that the present hydrological func-

115 tion is adjusted to the watershed geomorphology, we drew on insights from Litwin, Tucker,
116 et al. (2023) to hypothesize that:

- 117 1. Saturated areas and storm runoff are larger and more variable at Druids Run than
118 Baisman Run, and
- 119 2. Hillslopes are shorter and have less relief at Druids Run than Baisman Run.

120 First, we characterized the hydrological function and morphology of the two sites and
121 evaluated whether they support these hypotheses. To determine whether these differ-
122 ences could be the result of coevolution, we fully parameterized the landscape evolution
123 model used in Litwin, Tucker, et al. (2023) without calibration. We determined the im-
124 portance of subsurface hydrological differences by performing a simple sensitivity anal-
125 ysis in which we swapped the geomorphic process variables between the two sites and
126 observed whether geomorphic process rates could explain differences in emergent mor-
127 phology and hydrologic function.

128 2 Materials and Methods

129 2.1 Site descriptions

130 Our study sites are located in the Piedmont physiographic province, north of Bal-
131 timore, Maryland. The climate is humid, with a mean annual precipitation of approx-
132 imately 1150 mm and mean annual potential evapotranspiration of approximately 750
133 mm. There is no pronounced seasonality in precipitation, less than 5% falls of which falls
134 as snow. Baisman Run is a 381 ha watershed in Oregon Ridge Park, defined by an out-
135 let at (39.4795 N, 76.6779 W). Druids Run is a 107 ha watershed located in Soldiers De-
136 light Natural Environment Area, and is defined by an outlet at (39.4171 N, 76.8523 W).
137 The watersheds are 16 km apart, and are at approximately the same elevation (52 m and
138 56 m above sea level respectively). Both watersheds drain to the Chesapeake Bay; Bais-
139 man Run drains via the Gunpowder River and Druids Run via the Patapsco River. Bais-
140 man Run has been monitored extensively as part of the Baltimore Ecosystem Study, and
141 more recently as part of several projects aimed at improving understanding of deeply weath-
142 ered critical zones (Putnam, 2018; Cosans, 2022). Druids Run has no prior description
143 or study. It is unnamed in the National Hydrography Dataset, so we unofficially named
144 it in honor of a local group of druids that meet in the watershed.

145 Baisman Run is underlain by the Loch Raven Schist (Crowley et al., 1975), a Cambrian-
146 Devonian mica schist that has weathered to form deep, permeable soil and saprolite. Depth
147 to weathered bedrock is greater than 200 cm in most of the watershed, below saprolite
148 tens of meters thick at the ridge crests (Cosans, 2022). Above the saprolite, primary soils
149 include Manor loam and channery loam, Glenelg loam and channery loam, and Manor-
150 Brinklow complex in steeper slopes. Agriculture was historically present in the eastern
151 headwaters, where there is now suburban development, and a homestead and tree farm
152 were historically present in the Pond Branch sub-watershed (Cleaves et al., 1970). The
153 remainder of the watershed has been relatively undisturbed since the 1950s and today
154 supports a mature deciduous forest.

155 Druids Run is primarily underlain by the Soldiers Delight Ultramafite (Guice et
156 al., 2021). Soils are primarily classified as chrome silt loam, and are generally thin with
157 a strong permeability contrast at the base of the A horizon (at an average depth of 46
158 cm). Ridgetop soil is rocky and can be as thin as 5 cm, and exposed bedrock is common
159 near channel heads. In valley bottoms, alluvium and organic material accumulate to thick-
160 nesses around 1 m. The Soldiers Delight Ultramafite is host to a “serpentine barrens”
161 ecosystem, which consists primarily of grasses and shrubs with some areas supporting
162 hardwood and conifer trees. The Soldiers Delight area was mined for chromite in the 19th
163 and 20th century. Several small pits are present near ridge crests in Druids Run, and placer

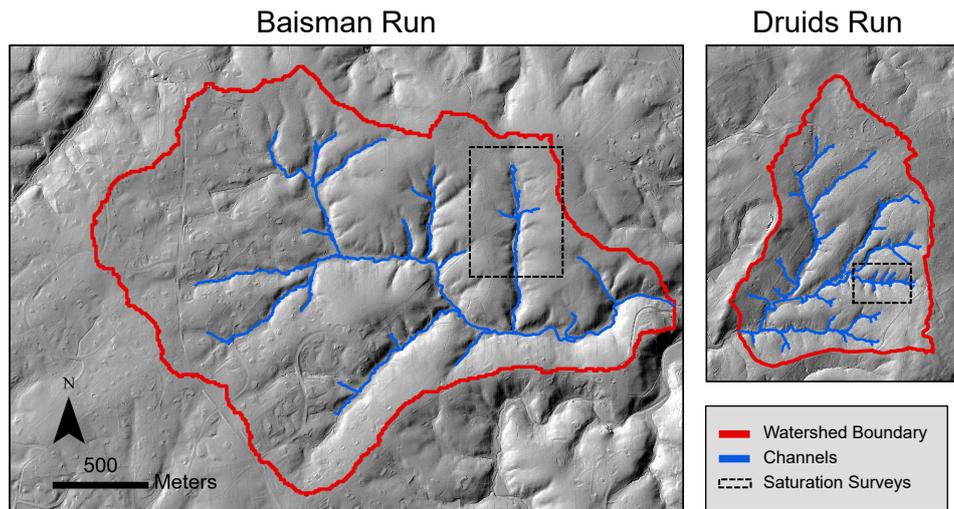


Figure 1. Hillshades of Baisman Run and Druids Run with the watershed boundary and channel network delineated with the DrEICH algorithm. Areas where we conducted saturation surveys (see Figure 4) are shown in dashed black boxes. The two sites are to scale, revealing the difference in their size and drainage dissection.

164 mining may have occurred in the valley bottoms, but the effects of this appear to be min-
 165 imal in this watershed. Some structures and two small ponds are present in the upper
 166 portion of Druids Run, but most of the watershed is free from development.

167 2.2 Hydrological data

168 We combined existing hydrological data with new measurements of precipitation,
 169 streamflow, and saturated areas. Instantaneous precipitation rates were measured from
 170 June 2022 to February 2023 at a weather station located in an open field approximately
 171 0.8 km north of Baisman Run. An identical unit was installed in an open area in Druids
 172 Run, which recorded instantaneous precipitation from April 2022 to February 2023. The
 173 stream gage at Baisman Run is operated and maintained by the U.S. Geological Survey
 174 (Gage 01583580). We established a new stream gage at Druids Run for this project.

175 The Druids Run stream gage is located at an existing concrete culvert crossing the
 176 stream channel. In April 2022 we installed a PVC housing on the concrete structure ap-
 177 proximately 2 m from the culvert inlet. We measured water stage with a Solinst Lev-
 178 elogger pressure transducer within that housing, and corrected for atmospheric pressure
 179 with a Solinst Barologger. The pressure transducer operated until the device failed in
 180 October 2022. Periodic discharge measurements were made to construct a rating curve.
 181 Low flows were measured with salt dilution gaging recorded with a HOBO conductiv-
 182 ity logger, and high flows were measured using an OTT MF Pro electromagnetic cur-
 183 rent profiler. A power law model fit the stage-discharge data well, as shown in Figure
 184 S1.

185 We surveyed limited areas of both watersheds manually for saturation conditions
 186 between April 2022 and March 2023. At Baisman Run, the surveys were conducted in
 187 the headwaters of the Pond Branch sub-watershed. At Druids Run, they were conducted
 188 in a headwater catchment near the eastern watershed boundary. We measured satura-
 189 tion at points along predefined transects, and returned to the approximate (but not ex-

act) positions for each survey. We selected transects to balance capturing a range of hill-slopes, zero- and first-order channels, while covering a small enough area to avoiding significant changes in saturation over the course of a measurement campaign. Saturation was measured by walking the transects, and pushing a rebar rod approximately 2 cm into the ground and moving the rod up and down in the shallow hole. Points along these transects were recorded as not saturated if no squishing sound was heard (N), soil-saturated if a squishing sound was heard (Ys), ponded (Yp), or flowing (Yf) if water was observed on the surface. Three close locations were measured at each point on the transect, and the highest category in this hierarchy was recorded as the value (e.g., if two points did not squish, but one did, the recorded class would still be Ys). This procedure was repeated under different discharge and moisture conditions.

2.3 Hydrological analysis

Valuable information about contributing areas can be extracted from rainfall and runoff timeseries. The event runoff ratio, defined as the ratio of the total event runoff to event precipitation, is an indicator of the proportion of the watershed that is contributing runoff during storms (e.g., O’Loughlin, 1986). To calculate event runoff, we first separated the discharge timeseries into baseflow and quickflow using the graphical approach described by Hewlett and Hibbert (1967). Baseflow is equal to discharge and quickflow is zero until discharge increases at a rate faster than $0.000546 \text{ m}^3 \text{ s}^{-1} \text{ km}^{-2} \text{ h}^{-1}$ (Hewlett & Hibbert, 1967). Baseflow continues to increase at this rate until discharge declines and is equal to baseflow. Storm events are periods where quickflow is greater than zero and the rise is associated with precipitation. We defined event precipitation as the total precipitation falling between a fixed time t_0 before the runoff event begins and a fixed time t_1 before the runoff event ends. By inspection of the timeseries, we found that $t_0 = 2$ hours and $t_1 = 1$ hour were appropriate for Druids Run, and $t_0 = 6$ hours and $t_1 = 2$ hours were appropriate for Baisman Run. We excluded runoff events shorter than 6 hours because these generally had small discharge responses relative to noise in the timeseries.

While the runoff ratio provides a signature of contributing area, the saturation dataset provides a direct means to assess the variability of saturated areas. The saturation surveys yielded categorical data that vary with topographic position and catchment discharge. To develop quantitative insights from the dataset, we first created a binary classification of whether points were not saturated (N) or saturated (Ys, Yp, Yf). We then used logistic regression to generalize our discrete measurements to predictions of how saturation probability varies with topographic (wetness) index (Beven & Kirkby, 1979) and discharge:

$$\log\left(\frac{p}{1-p}\right) = \alpha_0 + \alpha_1 \log\left(\frac{A}{v_0|\nabla z|}\right) + \alpha_2 \log\left(\frac{Q_b}{A_{tot}}\right) \quad (1)$$

where p is the probability of saturation, $TI = \frac{A}{v_0|\nabla z|}$ is the topographic index (note that here we do not include the log transform in the definition), Q is the discharge at the start of the saturation measurement campaign, A_{tot} is the watershed area, and α_0 , α_1 , and α_2 are model parameters.

2.4 Hillslope length and relief

We conducted geomorphic analyses using a lidar-derived digital elevation model with 0.76 m resolution, which was collected in 2015 and is publicly available from Baltimore County. We conducted all topographic analyses using LSDTopoTools (S. Mudd et al., 2022). To determine hillslope length and relief, we began by identifying the channel networks at both sites using the DrEICH algorithm (Clubb et al., 2014). DrEICH uses χ -analysis (Perron & Royden, 2013) to locate channel heads at the transition point from linear channel segments to nonlinear hillslope segments in χ -elevation space. χ -analysis is discussed in more detail in Section 2.7.2. We adjusted the DrEICH model parameters such that the predicted channel network matched the observed network in the subwa-

231 tersheds where we surveyed saturated areas. We then used the channel network to iden-
 232 tify hilltops, which are defined as edges shared by watersheds with the same Strahler stream
 233 order (Hurst et al., 2012). Finally, we calculated hillslope length as the steepest descent
 234 distance from each hilltop point to the nearest channel point, and hillslope relief as the
 235 hilltop elevation above the nearest channel point (Grieve et al., 2016).

236 2.5 Landscape evolution model

237 To test the link between hydrological and geomorphic features, we used the land-
 238 scape evolution model described in Litwin, Tucker, et al. (2023). The model accounts
 239 for topographic evolution due to baselevel change, water-driven erosion using the stream-
 240 power erosion equation, and hillslope sediment transport using a nonlinear hillslope dif-
 241 fusion equation. We decided to use a linear diffusion formulation, as the hillslopes at Bais-
 242 man Run and Druids Run generally remain convex until they reach valley bottoms, and
 243 the topography shows no evidence of shallow landsliding or other mass movements. The
 244 subsurface maintains constant and spatially uniform properties through evolution, im-
 245 plicitly assuming that the production of permeable material keeps pace with surface ero-
 246 sion. The overland flow that drives fluvial erosion is generated by exfiltration and pre-
 247 cipitation on saturated areas in places where the shallow aquifer reaches the land sur-
 248 face. The shallow aquifer model uses the Dupuit-Forcheimer assumptions to calculate
 249 flow over a sloping impermeable base. The aquifer receives recharge from the vadose zone,
 250 which is represented as a single 1-dimensional profile in which discrete depth increments
 251 fill and drain by the plant-available water capacity in the increment. Recharge is calcu-
 252 lated by determining the amount of water in the vadose profile that infiltrates below the
 253 water table depth at each point in the aquifer. The climate is treated as a simple ran-
 254 dom process, following Eagleson (1978), with exponentially distributed storm depth, du-
 255 ration, and interstorm duration, and constant evapotranspiration at the climatological
 256 mean rate during the interstorm periods.

257 We ran the model under the same initial and boundary conditions used in Litwin,
 258 Tucker, et al. (2023). The domain is square, and the bottom boundary is fixed to base-
 259 level, while the remaining three side boundaries are zero-flux. This allows for the estab-
 260 lishment of a drainage network with higher order streams than the same size domain where
 261 all boundaries are set to a fixed baselevel. In the absence of a known initial condition,
 262 we begin with a flat surface at baselevel. We ran the model for 50 Ma to approach dy-
 263 namic equilibrium between erosion and uplift. While this timescale is long relative to
 264 periodic changes in climate and baselevel in the Eastern Piedmont (e.g., Cleaves, 1989),
 265 we know that both sites have experienced the same forcings through their evolution, such
 266 that a single climate and baselevel change rate should still provide insights into their evo-
 267 lution.

268 2.6 Hydrological parameters

269 2.6.1 *Transmissivity, hydraulic conductivity, and permeable thickness*

270 The maximum transmissivity, which we will just call the transmissivity, is defined
 271 as the depth-integrated saturated hydraulic conductivity. It appears in our model as the
 272 product of the effective saturated hydraulic conductivity k_s and permeable thickness b .
 273 We developed a novel method to use the saturation survey data to estimate a catchment-
 274 averaged transmissivity, building on an existing approach. We then divided that value
 275 into estimates of k_s and b .

Our method of estimating transmissivity is similar to that described by O’Loughlin
 (1986), as it is built on a steady state hillslope water balance and the assumption that
 places with the same topographic index TI saturate at the same time (Beven & Kirkby,
 1979). The approach begins by considering recharge that is supplied at a rate $r(x, y)$ to

the saturated zone. At hydrologic steady state, the total water outflow along a topographic contour segment with length v_0 is equal to the integral of recharge over the area upslope of the contour A_c . The maximum amount of recharge that can be moved through the subsurface before saturation occurs depends on the transmissivity T and the local hydraulic gradient, which is assumed to be equal to the topographic gradient ∇z . As a result, the criterion for saturation is:

$$\int_{A_c} r(x, y) dA \geq T |\nabla z| v_0. \quad (2)$$

At saturation, any additional recharge will become overland flow. Because in general the recharge is not known, O'Loughlin (1986) equated the total watershed recharge with the watershed baseflow Q_b :

$$\int_{A_{tot}} r(x, y) dA = Q_b, \quad (3)$$

where A_{tot} is the watershed area. From this expression, we derived an average recharge rate $\bar{r} = Q_b/A_{tot}$. Dividing Equation 2 by the average recharge rate equation and rearranging the terms, we derived an expression for the discharge-normalized transmissivity:

$$\frac{1}{|\nabla z| v_0} \int_{A_c} \left(\frac{r}{\bar{r}} \right) dA \geq \frac{T}{Q_b/A_{tot}}. \quad (4)$$

By further assuming that the integral in the above expression is approximately unity, we found an expression that relates the topographic index to transmissivity and baseflow discharge:

$$\frac{A}{|\nabla z| v_0} \geq \frac{T}{Q_b/A_{tot}}. \quad (5)$$

We will call the topographic index where saturation begins to occur TI^* , which is a function of discharge Q_b . Using a log transform, we derived an expression for the log of transmissivity:

$$\log(T) = \log(TI^*) + \log\left(\frac{Q_b}{A_{tot}}\right). \quad (6)$$

To find T using this expression and our saturation surveys, consider a logistic regression model with the form:

$$\rho(p) = \log\left(\frac{p}{1-p}\right) = \beta_0 + \beta_1 \log\left(\frac{A}{v_0 |\nabla z|} \frac{Q_b}{A_{tot}}\right) \quad (7)$$

where β_0 and β_1 are parameters of the regression model. This logistic regression model is very similar to that in Equation 1, but has one fewer parameter, and consequently enforces that the odds of saturation are log-linearly dependent on the product of Q_b and TI . At the critical value of topographic index TI^* , we will call the odds of saturation ρ^* :

$$\rho^* = \beta_0 + \beta_1 \log\left(TI^* \frac{Q_b}{A_{tot}}\right). \quad (8)$$

Finally, we rearranged Equation 8 to match the form of Equation 6, and solved for the transmissivity:

$$T = e^{(\rho^* - \beta_0)/\beta_1}. \quad (9)$$

276 The main difference between this approach and that described by O'Loughlin (1986) is
 277 that their approach equates the event runoff ratio with the proportion of the watershed
 278 that is saturated, while we have direct estimates of the saturated area. This should make
 279 our approach more robust, though it is still limited to the steady-state theory from which
 280 it was derived. Finally, we partitioned transmissivity between permeable thickness b and
 281 an effective saturated hydraulic conductivity k_s based on permeable thickness values taken
 282 from the USDA Soil Survey (Staff & Natural Resources Conservation Service, United
 283 States Department of Agriculture., 2023) and insights gained from prior subsurface in-
 284 vestigations of Baisman Run.

2.6.2 Drainable porosity and plant available water content

Drainable porosity n_e relates the depth of water stored or released to the change in hydraulic head. Estimates usually require either hydraulic well tests or laboratory analyses. In the absence of hydraulic test data or permission to take soil samples from Druids Run, we assumed that the drainable porosity was the same at both sites. Plant available water content (n_a) is the amount of water, below the field capacity, that is available for plant use. The values were estimated based on the USDA Soil Survey data for the dominant soil types at the two sites.

2.6.3 Climatological parameters

We fit three independent exponential distributions for storm depth d_s , duration t_r , and interstorm duration t_b by calculating the mean values of these quantities from a precipitation dataset previously collected from 2014-2018 at the weather station at Baisman Run (Cosans, 2022). Because the two sites are very close together, this one time-series was used to calculate storm statistics at both sites. The distributions are:

$$f(d_s) = \frac{1}{\langle d_s \rangle} \exp\left(-\frac{d_s}{\langle d_s \rangle}\right) \quad (10)$$

$$f(t_r) = \frac{1}{\langle t_r \rangle} \exp\left(-\frac{t_r}{\langle t_r \rangle}\right) \quad (11)$$

$$f(t_b) = \frac{1}{\langle t_b \rangle} \exp\left(-\frac{t_b}{\langle t_b \rangle}\right) \quad (12)$$

$$(13)$$

where the angled braces indicate the temporal mean of the quantity. Potential evapotranspiration (ET) was estimated based on the average annual value in Baltimore between 1981 and 2010, as reported by the Northeast Regional Climate Center at Cornell University. In our model, ET only occurs during interstorm periods, so the interstorm potential ET rate pet was estimated by rescaling the average potential ET rate with the interstorm time fraction. Our climatological approach is simplistic, neglecting covariance of storm depth, duration, and interstorm duration, seasonality, paleoclimatic variability, and so on. However, we do not expect any large differences in the climate between the two sites, so even a simple approach should allow us to make comparisons of how landscapes with different geomorphic and subsurface hydrologic properties respond to climatic conditions similar to those observed at our sites.

2.7 Estimating geomorphic parameters

The topographic parameters of our model are the uplift or baselevel change rate U , hillslope diffusivity D , streampower incision coefficient K , characteristic contour width v_0 , and the streampower exponents m and n , as discussed below. The Piedmont is thought to be in geomorphic steady state (Pavich, 1989; Bazilevskaya et al., 2013), so the regional rate of baselevel change was estimated the long-term erosion rate estimated with cosmogenic ^{10}Be . The remaining parameters were identified using topographic analysis.

2.7.1 Hillslope diffusivity

Hillslope diffusivity can be derived from the rate of baselevel change U and hill-top curvature C_{HT} (Roering et al., 2007; Hurst et al., 2012):

$$D = -\frac{U}{C_{HT}}. \quad (14)$$

In hillslope evolution contexts, it is typical to account for the ratio of the bulk densities of regolith (on which the diffusion process occurs) and parent material (on which baselevel change occurs) (Roering et al., 2007). Because we are working with an integrated

316 channel and hillslope model, and we do not have good estimates for the bulk density of
 317 fluviially-eroded material, we will neglect the bulk density terms. In this context, D is
 318 an effective diffusivity that will match the simulated hilltop curvature with that from
 319 our topographic measurements. We calculated the hilltop curvature by taking the sec-
 320 ond derivative of a polynomial surface fit to a 10 m footprint around each hilltop point.
 321 Hilltop points are the same as those used for the hillslope length analysis. The footprint
 322 size was selected by calculating the hilltop curvature for footprints of varying sizes and
 323 selecting the size at which there is a break in the standard deviation of curvature, fol-
 324 lowing the procedure described by Hurst et al. (2012).

325 2.7.2 Streampower parameters

We estimated the streampower law parameters using an integral approach called χ -analysis (Perron & Royden, 2013). While the parameters can be derived from slope-area analysis, slope estimates often have significant noise that can result in poor parameter estimates (Perron & Royden, 2013). The integral approach is more stable, as it only requires the elevation and the upslope area to calculate the model parameters. The typical χ -analysis needed slight modification to accommodate our landscape evolution model. Litwin et al. (2022) derived the fluvial incision term of the landscape evolution model with assumptions that yielded linear dependence on the dimensionless discharge Q^* , a slope exponent $n = 1$, and area exponent $m = 1/2$. We derived a more general form by assuming that the exponent that determines the channel width from area and the exponent that determines erosion rate from shear stress were free parameters:

$$E_f = KQ^{*n} (v_0a)^m |\nabla z|^n \quad (15)$$

where E_f is the fluvial incision rate, K is the erodibility, v_0 is the characteristic contour width, a is the area per contour width, and ∇z is the elevation gradient. For simplicity, we will use the variable Q^* to refer to the temporally-averaged dimensionless discharge which is called $\langle Q^* \rangle$ in Litwin, Tucker, et al. (2023). Because χ -analysis is usually only applied to river channels, it is typical to neglect the hillslope diffusion term, and write the solution at equilibrium between uplift and fluvial incision along a channel distance coordinate x :

$$U = KQ^{*n} (v_0a)^m \left| \frac{\partial z}{\partial x} \right|^n. \quad (16)$$

We then solved for $|\partial z/\partial x|$, and substituted area for area per contour width times the characteristic contour width $A = v_0a$:

$$\left| \frac{\partial z}{\partial x} \right| = \left(\frac{U}{KQ^{*n}} \right)^{1/n} A^{-m/n}. \quad (17)$$

Next we normalized upslope area to a reference drainage area A_0 , and integrate the equation above with respect to x :

$$z(x) = z(x_b) + \int_{x_b}^x \left(\frac{U}{KQ^{*n} A_0^m} \right)^{1/n} \left(\frac{A_0}{A(x)} \right)^{m/n} dx \quad (18)$$

where $z(x_b)$ is the elevation at a specified baselevel location x_b . In general, Q^* varies with position, so we cannot remove it from the integral. However, in our model Q^* generally approaches a constant value with distance downstream equal to one minus the actual evapotranspiration relative to precipitation $1 - \langle AET \rangle / \langle P \rangle$, which is approximately the mean runoff ratio $\langle Q \rangle / \langle P \rangle$. We will call this value Q_{max}^* . Then we can write:

$$z(x) = z(x_b) + \left(\frac{U}{KQ_{max}^{*n} A_0^m} \right)^{1/n} \chi, \quad (19)$$

where

$$\chi = \int_{x_b}^x \left(\frac{A(x)}{A_0} \right)^{m/n} dx. \quad (20)$$

These equations show that the elevation of a stream channel in dynamic equilibrium should be linear with respect to χ if U , K , and Q_{max}^* are uniform, and that the slope of that relationship should be:

$$k_{sn} = \left(\frac{U}{K Q_{max}^* A_0^m} \right)^{1/n}, \quad (21)$$

326 which is often called the normalized channel steepness index. Note that this is related
327 to but distinct from our use of “steepness” in Litwin et al. (2022).

We calculated the slopes of channel segments in χ -elevation space for the channel networks we extracted previously. Because the reference drainage area A_0 is introduced for dimensional purposes only, we can set it equal to unity, and solve for the streampower incision coefficient K :

$$K = \frac{U}{(k_{sn} Q_{max}^*)^n}. \quad (22)$$

328 3 Results

329 3.1 Hydrologic and geomorphic observations

330 3.1.1 Discharge, baseflow, and runoff ratio

331 Figure 2 shows the timeseries of discharge and precipitation for both sites. Base-
332 flow (in dark blue) at Baisman Run shows a strong annual signal, with drydown from
333 early summer continuing until October, when a small persistent increase is combined with
334 episodic increases in response to large storms. Unfortunately the discharge timeseries avail-
335 able to us at Druids Run is too short to look at annual trends, though there does ap-
336 pear to be a significant drydown from spring into summer, leading to low flows by late
337 June. We did not observe no-flow conditions at the gage location, but we do know that
338 flows were often close to or below the pressure transducer detection limit during the sum-
339 mer.

340 The storm runoff ratio is substantially more variable at Druids Run than Baisman
341 Run. We identified 21 storm events at Druids Run and 43 storm events at Baisman Run,
342 and found that the total event precipitation explained most of the variation in total event
343 quickflow $Q_{f,event}$ (Figure 3). Events are colored by the antecedent baseflow, which shows
344 that some of the variation in event runoff that cannot be explained by event precipita-
345 tion may be explained by antecedent conditions. To quantify the sensitivity of event runoff
346 to event precipitation, we fit the curve $Q_{f,event} = a_2 P_{event}^{a_1}$, where the log-space slope
347 corresponds to the fitted exponent a_1 . The exponent and standard error are 3.17 ± 0.40
348 and 1.89 ± 0.13 at Druid Run and Baisman Run, respectively. An exponent $a_1 = 1$ would
349 indicate that the storm runoff is a constant proportion of the event precipitation. When
350 the event runoff ratio is interpreted as the effective proportion of the watershed contribut-
351 ing runoff (O’Loughlin, 1986), an exponent closer to one indicates that the contribut-
352 ing area does not vary with storm size. This interpretation suggests that contributing
353 areas vary with precipitation at both sites, but they are more variable at Druids Run
354 than Baisman Run. This interpretation also suggests that as storm events approach 100
355 mm, nearly all of Druids Run contributes storm runoff (3A). These events are fairly fre-
356 quent; the annual maximum recurrence interval of 100 mm of precipitation in 24 hours
357 is approximately two years at our sites (NOAA, 2024).

358 3.1.2 Saturated areas

359 At Druids Run, observed saturation was highly variable in time and correlated with
360 discharge. We measured saturation five times along nine transects, seven of which run
361 along first order drainages or the interflaves between them, and two of which run par-
362 allel to the valley bottom (Figure 4A–E). The surveys conducted under the two high-
363 est flow conditions (C, E) had the greatest number of saturated points. Saturation was

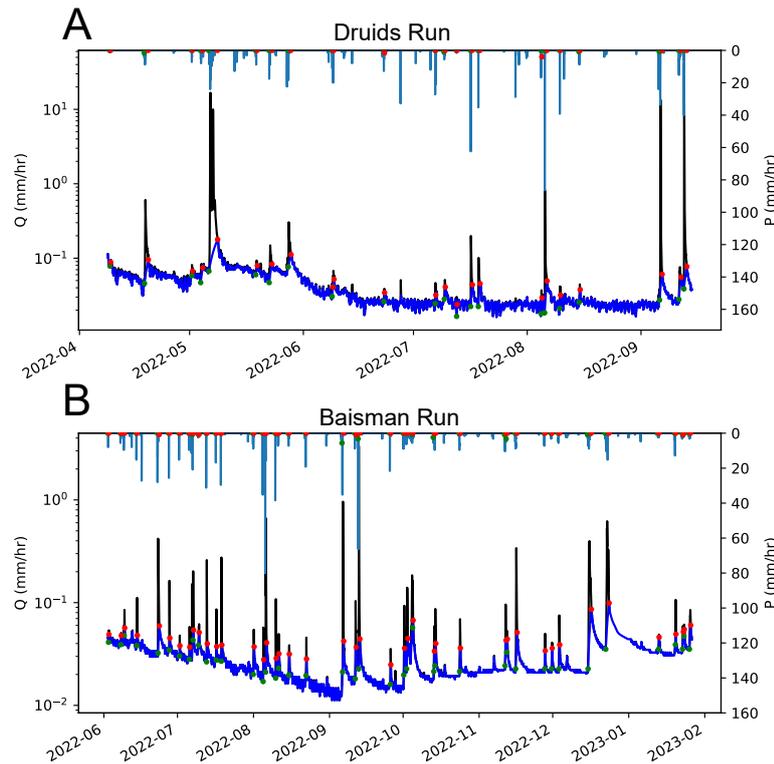


Figure 2. Timeseries of discharge Q (black), baseflow Q_b (dark blue), and precipitation P (light blue) at Druids Run (A) and Baisman Run (B). Storm events that we identified based upon the baseflow separation and precipitation begin with green dots and end with red dots, which are placed at the corresponding times on both the precipitation and discharge timeseries. Note that the timeseries for Baisman Run and Druids Run are not aligned in time.

364 often discontinuous with distance downstream in first order channels. Upslope areas some-
 365 times saturated and flowed first, while downslope reaches remained dry, as flow passed
 366 through the subsurface. First order channels tend to have exposed bedrock or thin al-
 367 luvial cover near their headwaters, while closer to the valley bottom they become sub-
 368 merged in alluvium that has sufficient capacity to move the water from upslope through
 369 the subsurface. Bedrock fractures may also play a role in redistributing surface flow to
 370 subsurface pathways.

371 In contrast, saturated areas were more static at Baisman Run. We measured satura-
 372 tion four times along six transects, four of which run perpendicular to the valley bot-
 373 tom, and two run parallel to it (Figure 4F–I). Regardless of discharge, we found that satura-
 374 tion was confined to locations at or below the distinct break in slope where the hill-
 375 slopes meet the valley bottom. Within the valley bottom, saturation was not present ev-
 376 erywhere, as the stream channel is incised into the valley bottom alluvium in some places.
 377 Flow emerges at distinct springs and seeps at the break in slope (Putnam, 2018). The
 378 springs are further evidence that subsurface pathways support baseflow, while the rela-
 379 tively static nature of saturated areas support our observation that event quickflow is
 380 less sensitive to event precipitation at Baisman Run than it is at Druids Run.

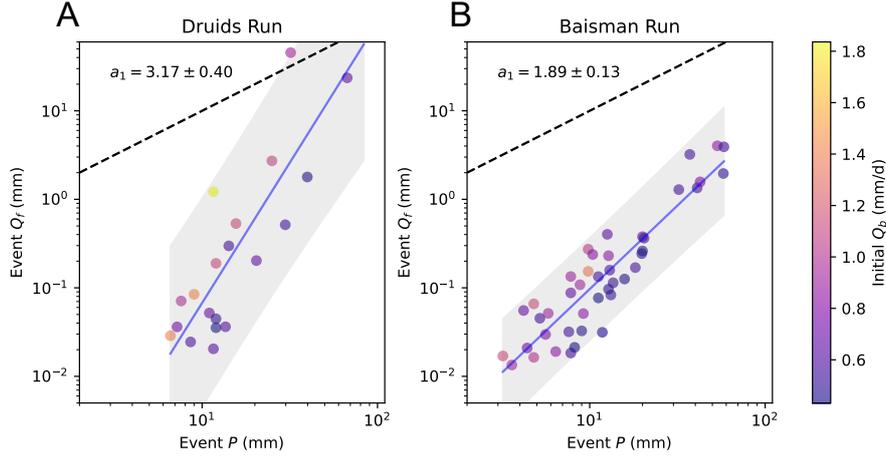


Figure 3. Event runoff characteristics for Druids Run (A) and Baisman Run (B). Event totals are calculated by summing the 15-minute precipitation and quickflow timeseries over the event durations. The points are colored by the initial baseflow Q_b . The dotted line is a 1:1 line, which represents the case where event runoff is equal to event precipitation. The blue line is a power law regression with the form $Q_{f,event} = a_2 P_{event}^{a_1}$, and the shaded area is the 95% confidence interval on the regression. The range on the coefficient a_1 is given as the standard error.

381 We generalized our point observations to whole-watershed predictions of saturation
 382 frequency using logistic regression. Specifically, we predicted the presence of sat-
 383 uration (flowing water, ponded water, or soil saturation) using topographic index and
 384 discharge using Equation 1. The parameters of the fitted model are shown in Table 1.
 385 To calculate topographic index, we first resampled the DEM to 5 m resolution to smooth
 386 over roughness in the high resolution DEM and to reflect the uncertainty in the posi-
 387 tioning data of our saturation surveys. The resampling approach is also consistent with
 388 our measurement scheme, in which we labeled locations based upon the highest sat-
 389 uration class observed in a small vicinity. We calculated upslope area using the D_∞ algo-
 390 rithm, and slope using the same 10 m footprint used to calculate hilltop curvature. While
 391 our regression model calls for the use of baseflow discharge, we used the total discharge,
 392 as all of our samples were taken during baseflow or recession periods. This was also nec-
 393 ecessary because the timeseries of discharge at Druids Run does not overlap all the sat-
 394 uration surveys. For consistency, we used instantaneous discharge measurements from
 395 immediately before the surveys began. At Druids Run, we made these measurements us-
 396 ing dilution gaging; Baisman Run, we used instantaneous discharge from the USGS gage.

	α_0	α_1	α_2
Druids Run	4.609 ± 0.637	0.174 ± 0.040	1.000 ± 0.097
Baisman Run	-7.487 ± 3.922	0.703 ± 0.103	0.081 ± 0.540

Table 1. Estimated parameter values of the logistic regression models for saturation (Equation 1), where α_0 is the intercept, α_1 is the coefficient on topographic index, and α_2 is the coefficient on the area-normalized discharge. Parameter ranges are given as standard errors.

397 We used the logistic model to predict the odds of saturation for the range of to-
 398 pographic index values in each watershed and the range of discharge values at which sat-

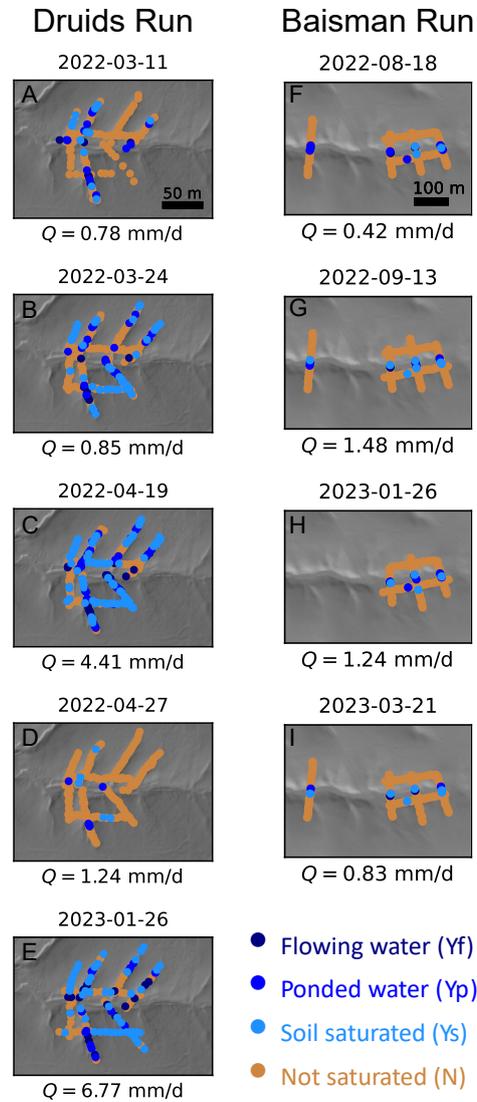


Figure 4. Observations of saturation made on transects at Druids Run (A–E) and Baisman Run (F–I). The latter plots have been rotated 90 degrees such that north is in the direction of the positive x-axis. In both figures, flow in the valley bottom is from right to left. The classification and sampling approaches are described in Section 2.2.

uration surveys were conducted (Figure 5). In Figure 5, the topographic index value at which the black dashed line intersects the odds ratio curves is the critical value of TI where saturation becomes more likely than not for a given value of discharge. We plotted this together with the probability density of watershed topographic index (orange) to show how the critical TI relates to the distribution of TI for the watershed.

The regression model for Druids Run in Figure 5A shows that the predicted odds of saturation varies substantially with discharge. When discharge is small, the critical TI value confines likely saturation to a very small portion of the total watershed area, while for large discharge the critical value of TI is low enough that most of the watershed is likely to be saturated. This supports the high variability of saturation in space and time that we inferred from the pointwise measurements.

The logistic regression model predicts very different behavior for Baisman Run (Figure 5B). First, we notice that the saturation odds curve does not vary with discharge, such that all curves overlap. This is reflected in the regression parameter α_2 on discharge (Table 1), which is much smaller and more uncertain for Baisman Run than Druids Run. As a result, the critical value of topographic index is nearly constant with time. Second, we notice that the curves are narrower and steeper than those estimated for Druids Run, such that the odds of saturation increases more abruptly around the critical value of TI . This is reflected in the regression parameter α_1 on topographic index, which is much larger at Baisman Run than Druids Run. This supports our observation that saturation emerges abruptly at the transition from hillslopes to valley bottoms.

The logistic regression models also allowed us to generalize the saturation predictions to the entire watersheds. We predicted saturation through time for the discharge timeseries in Figure 2 and for all raster points based upon their topographic index. We then classified whether each point was “wet” (exceeded criteria for saturation greater than 95% of the time), “dry” (exceeded criteria for saturation less than 5% of the time), or variably saturated if it met neither of those criteria.

Figure 6 shows a dramatic difference in the hydrological function of the two sites based on the logistic regression model predictions. The predicted channel network at Druids Run was ephemeral until close to the watershed outlet. Saturation occurred occasionally in zero-order basins and up onto the hillslopes. Some of the hillslopes we sampled that appear as “dry” may in fact saturate occasionally, but less than 5% of the time. In contrast, the regression model predicted that Baisman Run had a continually wet stream channel over the course of our observation period, and did not experience saturation on the hillslopes.

Analysis of rainfall-runoff and saturation data reveal the dramatic difference between hydrological function of the two sites. When the permeable subsurface is thin, as at Druids Run, much of the landscape saturates and desaturates relatively easily in response to precipitation, and the effective proportion of the watershed contributing runoff varies substantially. In contrast, when the permeable subsurface is thick, as at Baisman Run, the same precipitation causes modest or no change in saturated areas, though new subsurface flow paths may still be activated with increasing storm size, such that the effective contributing area increases with increasing wetness.

3.1.3 Hillslope length and relief

Both hillslope length and relief are greater at Baisman Run than Druids Run. The channel networks and hilltop points from which hillslope length and relief were defined are shown in Figure 1. Totals of 5.3×10^4 and 7.0×10^4 hilltop points with unique length and relief were identified at Druids Run and Baisman Run, respectively. The median hillslope length is 88.3 m at Druids Run and 177.3 m at Baisman Run, while median relief was 2.9 m at Druids Run and 6.7 m at Baisman Run. Figure 7A shows that there is no

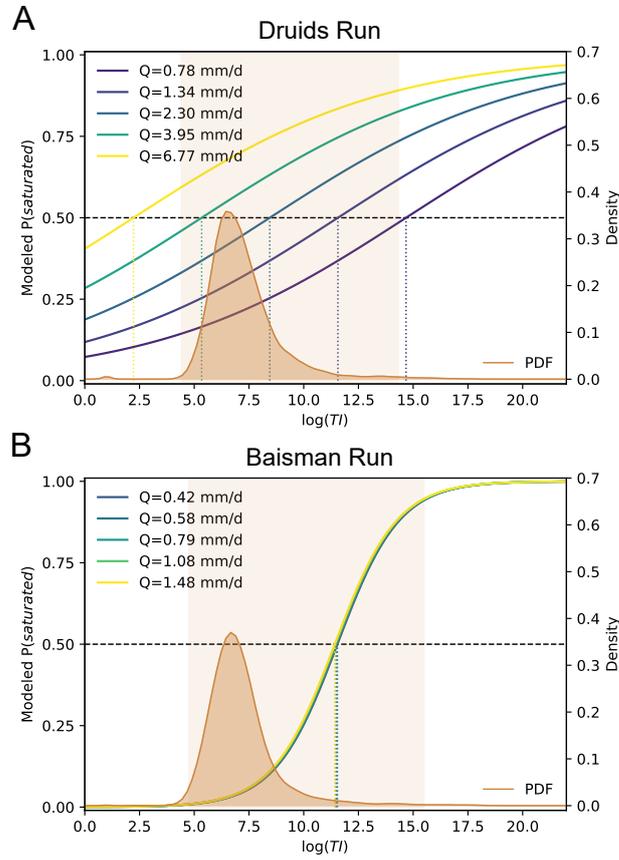


Figure 5. Regression results for Druids Run (A) and Baisman Run (B). The regression model has the form given in Equation 1. The modeled probability of saturation is given in terms of topographic index and discharge, where discharge varies logarithmically across the range of saturation survey discharge values. There is a dashed line at the 50% probability mark, and where this intersects each one of the probability curves, there is a dotted line dropped to the x-axis. This indicates the critical value of topographic index at which saturation is more likely than not to occur given that value of discharge. On the opposing axis is the probability density of topographic index, estimated with a kernel density approach. The lighter shaded region indicates the range of TI values sampled in our surveys, which indicates good topographic index coverage of our samples.

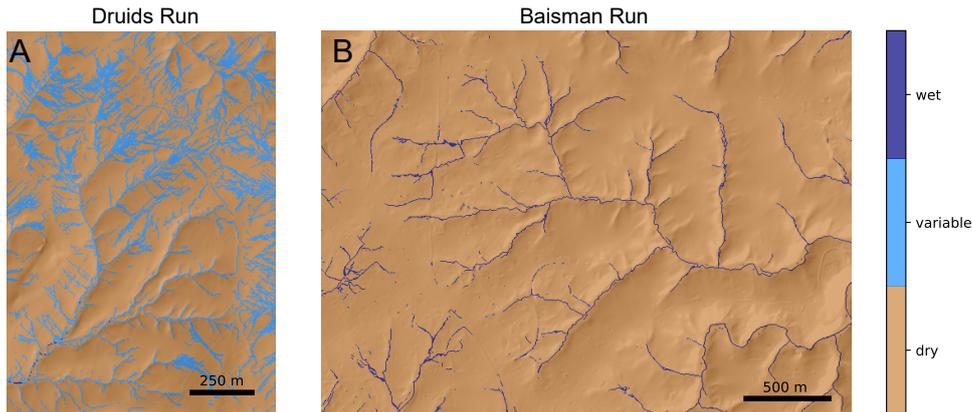


Figure 6. Classified saturated areas for Druids Run (A) and Baisman Run (B), based on the logistic regression model in Equation 1 and shown in Figure 5, and the runoff timeseries shown in Figure 2. The modeled probability necessary for saturation was set at 50%. A location was classified as “wet” if it exceeded criteria for saturation greater than 95% of the time, “dry” if it exceeded criteria for saturation less than 5% of the time, or variably saturated if it was in between. At Druids Run we predicted persistent saturation near the watershed outlet, an ephemeral channel network above that, and occasional saturation on some flat and concave hillslopes, and generally dry convex hillslopes. At Baisman Run, we predicted persistent saturation in the channel network, and dry conditions everywhere else.

449 overlap in the interquartile range (IQR) of hillslope length or relief for the two sites. The
 450 strength and sign of this difference supports our hypothesis that the site with a thick per-
 451 meable subsurface will have greater hillslope length and relief than that with a thin per-
 452 meable subsurface.

453 3.2 Landscape evolution parameterization

454 While both the hydrological and geomorphic differences between Druids Run and
 455 Baisman Run support our hypotheses, we have not yet established that the subsurface
 456 is the link between the emergent hydrological function and morphology. To do so, we es-
 457 timated the parameters for DupuitLEM, and ran the model under conditions that ap-
 458 proximate those found at our sites. Using the approaches described in Sections 2.6 and
 459 2.7, we estimated all the parameters needed to run the model without calibration.

460 3.2.1 Hydrologic parameters

461 We first estimated the transmissivity using Equation 9. We estimated the param-
 462 eters β_0 and β_1 by fitting Equation 7 using topographic index, discharge, and saturation
 463 survey data. With the fitted model, we determined the optimal threshold probability p^*
 464 at which saturation was likely to occur. While we could have chosen 50% as we did in
 465 the regression model for saturated area, we found that this performed poorly on the sim-
 466 pler two-parameter formulation used to calculate transmissivity. The selected value of
 467 p^* should balance correctly classifying points as saturated (high true positive ratio (TPR))
 468 and minimizing the number of points that are misclassified as saturated (low false pos-
 469 itive ratio (FPR)). Plotting TPR against FPR gives the receiver operating character-

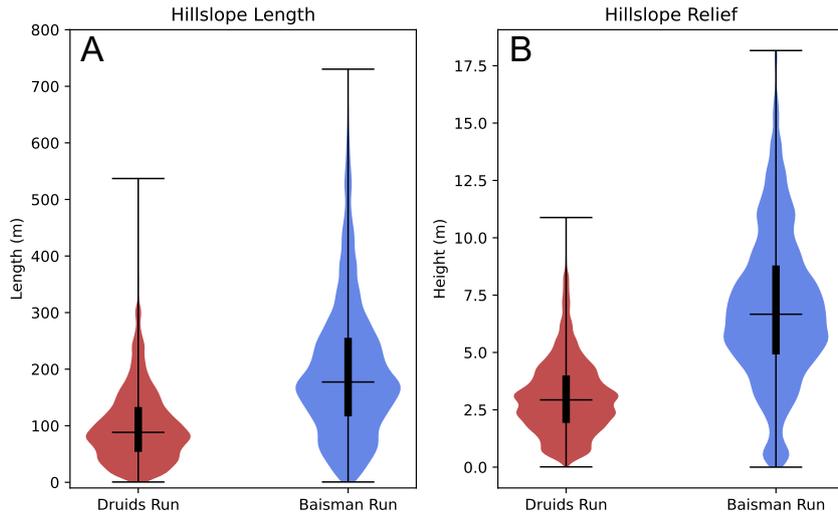


Figure 7. Violin plots of hillslope length and relief for Druids Run (A) and Baisman Run (B). Hillslope length is the length along a flow path from a hilltop point to the nearest channel point along a flowpath. Hillslope relief is the drop in elevation over that distance. Violin plots show the median, minimum, and maximum (horizontal lines) values and the interquartile range (wider vertical bar).

470 istic curve, from which we selected the optimal threshold probability by maximizing the
 471 difference TPR-FPR. The results of this process are shown in Figure 8. Using the op-
 472 timal p^* , we estimated the transmissivity from Equation 9 10,000 times using Monte Carlo
 473 simulations to determine the uncertainty due to the variance and covariance of the lo-
 474 gistic regression parameters. The median and quartiles of transmissivity are reported in
 475 Table 2. This approach predicts that the transmissivity at Baisman Run is nearly 8 times
 476 higher than at Druids Run. There is no overlap between the IQRs of the estimated trans-
 477 missivities, which suggests a robust difference between the two sites. While the true un-
 478 certainty is likely much larger as a result of methodological choices (raster resolution,
 479 flow routing method, threshold selection method), experimentation suggested that the
 480 median transmissivity is always larger at Baisman Run than Druids Run when the same
 481 methodology is applied to both sites.

	Transmissivity (m^2/d)			Regression Parameters			
	Med	LQ	UQ	$\bar{\beta}_0$	$\bar{\beta}_1$	ρ^*	p^*
Druids Run	1.12	0.88	1.40	-0.691	0.268	-0.660	0.341
Baisman Run	8.46	7.07	10.23	-3.113	0.676	-1.668	0.159

Table 2. Median (Med), lower and upper quartiles (LQ, UQ) of transmissivity estimated from the logistic regression model, and the associated regression model parameters. The bar over a variable indicates the mean value.

482 To estimate the effective hydraulic conductivity from transmissivity, we first esti-
 483 mated the permeable thickness. At Druids Run, data from the USDA Soil Survey sug-

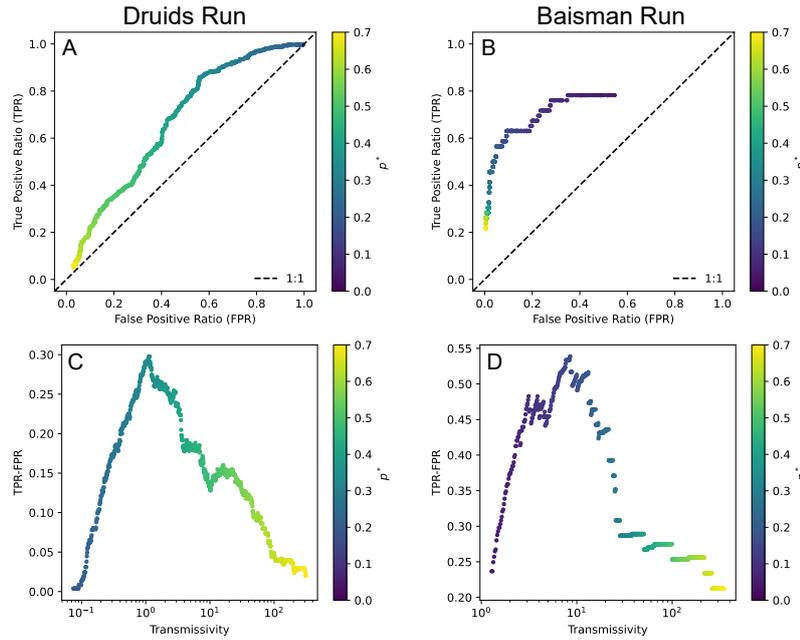


Figure 8. Results of the TPR-FPR analysis. (A–B) The receiver operating characteristic curve for Druids Run and Baisman Run, respectively, colored by the threshold value p^* used to obtain each combination of quantities. (C–D) The difference TPR-FPR, which we seek to maximize, plotted against the transmissivity value associated with each threshold p^* . We selected the transmissivity associated with the largest value of TPR-FPR.

484 gested a strong permeability contrast at the base of the A horizon, so we used the char-
 485 acteristic A horizon thickness as our permeable thickness (Staff & Natural Resources Con-
 486 servation Service, United States Department of Agriculture., 2023). At Baisman Run,
 487 there is no strong permeability contrast within the soil profile, so we used the entire soil
 488 profile thickness, weighted for the different soil types found in the watershed. We added
 489 2 m to this value to account for the importance of flow through the shallow saprolite (Cosans,
 490 2022), which is below the maximum depth considered by the USDA Soil Survey. We di-
 491 vided transmissivity by the permeable thickness, and found that the effective hydraulic
 492 conductivity is similar between the two sites (2.83×10^{-5} and 2.43×10^{-5} for Druids
 493 Run and Baisman Run, respectively); the majority of the difference in transmissivity is
 494 due to the difference in permeable thickness. The values are shown in Table 3.

495 We estimated drainable porosity and plant-available water content from literature
 496 values. We assumed drainable porosity was constant and equal to 0.25 at both sites, which
 497 is typical for materials with medium sand to medium gravel texture (Johnson, 1967). While
 498 drainable porosity is an important variable for regulating the degree to which the wa-
 499 ter table rises and falls in response to recharge, it has a relatively narrow range of pos-
 500 sible values in comparison to other parameters, so a possible difference between the sites
 501 should not have a strong effect on our results. We estimated plant-available water con-
 502 tent as 0.19 and 0.14 for Druids Run and Baisman Run respectively using characteris-
 503 tic values for our sites from the USDA Soil Survey.

504 Lastly, climatological variables were estimated using the approaches described in
 505 the methods section with weather station data and literature values. The relevant val-
 506 ues are shown in Table 3.

507 Overall, our characterization of hydrological properties is a rather coarse simpli-
 508 fication of reality. Exponential distributions for precipitation do not capture the impor-
 509 tance of extreme events (Rossi et al., 2016), while subsurface properties are often dynamic
 510 in time and vary both with depth and landscape position (e.g., St. Clair et al., 2015; Pe-
 511 drazas et al., 2021). Still, our prior modeling work showed that even with these simpli-
 512 fications the model can produce rich and complex emergent hydrologic behavior (Litwin,
 513 Tucker, et al., 2023). Our approach here can serve as a starting place for future work that
 514 accounts for higher-order controls on runoff generation.

Name	Symbol	Units	Druids Run	Baisman Run
Hydraulic conductivity	k_s	m/s	2.84e-5	2.43e-5
Permeable thickness	b	m	0.46	4.03
Plant-available water content	n_a	-	0.19	0.14
Drainable porosity	n_e	-	0.25	0.25
Mean storm duration	$\langle t_r \rangle$	s	1.02e4	1.02e4
Mean interstorm duration	$\langle t_b \rangle$	s	1.11e5	1.11e5
Mean storm depth	$\langle d_s \rangle$	m	4.50e-3	4.50e-3
Interstorm potential ET rate	pet	m/s	2.58e-8	2.58e-8

Table 3. All hydrological parameters needed to run DupuitLEM. The values for n_e , $\langle t_r \rangle$, $\langle t_b \rangle$, $\langle d_s \rangle$, and pet are identical at the two sites.

515 3.2.2 Geomorphic parameters

516 The uplift or baselevel change rate U is an important model parameter and is needed
 517 to obtain estimates of both the hillslope diffusivity D and the streampower incision co-
 518 efficient K . We equated U with the denudation rate estimated from in-situ ^{10}Be , assum-
 519 ing that the Piedmont physiographic province is near dynamic equilibrium between base-
 520 level change and denudation. Portenga et al. (2019) estimated the mean denudation rate
 521 of the Piedmont in the nearby Potomac River basin as 11.4 m/Myr (IQR 7.6 – 15.0) as-
 522 suming an average rock density of 2700 kg/m³. To quantify the uncertainty in U , and
 523 its contribution to the uncertainty in D and K , we estimated a probability distribution
 524 for U based on the box plot in Figure 4 of Portenga et al. (2019). The data did not ap-
 525 pear particularly skewed, so we modeled denudation with a normal distribution, which
 526 we truncated to permit only positive values.

527 We estimated the diffusivity based on hilltop curvature, as presented in Equation
 528 14. All the parameter values needed are shown in Table 4, and the distributions of the
 529 log of hilltop curvature are shown in Figure 9A. Hilltop curvature is quite similar at both
 530 sites. This is surprising since different processes likely contribute to diffusive transport
 531 at Druids Run versus Baisman Run. For example, freeze-thaw effects may be more im-
 532 portant in the exposed, rocky soils at Druids Run, while treethrow may be more impor-
 533 tant in the forest-covered soils at Baisman Run. We estimated the diffusivity and its un-
 534 certainty by Monte Carlo simulation, sampling the distribution of U 10,000 times, and
 535 selecting 10,000 values from the hilltop curvature dataset independently with replace-
 536 ment. The distributions of diffusivity from the Monte Carlo simulation are shown in Fig-
 537 ure 9B. The median diffusivity is 8.6e-3 m²/yr (IQR 4.4e-3 – 1.7e-2) at Druids Run,
 538 and 9.3e-3 m²/yr (IQR 4.3e-3 – 1.9e-2) at Baisman Run.

	C_{HT} (m^{-1})			U (m/yr)		
	Med	LQ	UQ	Med	LQ	UQ
Druids Run	$-1.272\text{e-}3$	$-2.084\text{e-}3$	$-7.053\text{e-}4$	$1.193\text{e-}5$	$7.561\text{e-}6$	$1.495\text{e-}5$
Baisman Run	$-1.125\text{e-}3$	$-2.123\text{e-}3$	$-6.571\text{e-}4$	$1.193\text{e-}5$	$7.561\text{e-}6$	$1.495\text{e-}5$

Table 4. Hilltop curvature C_{HT} and uplift U for Baisman Run and Druids Run. Negative curvature indicates convexity. Uplift values are the same for both sites.

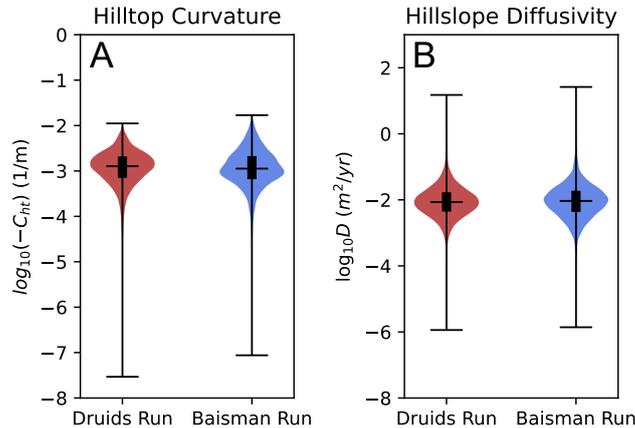


Figure 9. Violin plots of the log of hilltop curvature and log of hillslope diffusivity for Druids Run (A) and Baisman Run (B). Violin plots show the median, minimum, and maximum (horizontal lines) values and the interquartile range (wider vertical bar). Both distributions are similar, though Druids Run has slightly higher curvature, and therefore slightly lower diffusivity.

539 We calculated the streampower incision coefficient K using Equation 22 by esti-
540 mating n , k_{sn} , and Q_{max}^* . We first conducted a χ -analysis of the channel networks of
541 both sites to determine the streampower exponent n and then the appropriate steepness
542 index k_{sn} . Lastly, we estimated the maximum dimensionless discharge Q_{max}^* based on
543 available hydrologic data.

544 To calculate the optimal coordinate χ , we need to estimate the concavity index m/n
545 (see Equation 20) for which the channel network collapses to a single line in χ -elevation
546 space (Perron & Royden, 2013). We tried a range of values for the concavity index and
547 determined that $m/n = 1/2$ produced a satisfactory collinearity of channels for both
548 of the sites. Independently estimating the exponents m and n is challenging (Harel et
549 al., 2016), so we chose the combination $m = 1/2$ and $n = 1$ for consistency with our
550 prior modeling studies.

551 We determined k_{sn} from the slope of the relationship between χ and elevation for
552 individual channel segments using the method described by S. M. Mudd et al. (2014).
553 We estimated K using the segments that are above the 40th percentile of channel net-
554 work drainage area, which are colored by k_{sn} in Figure 10A–B. We selected this drainage
555 area cutoff to isolate channel segments where Q^* is less likely to vary with distance down-
556 stream. We found that channel segments with smaller upslope areas were often less lin-
557 ear in χ -elevation space, which may indicate a change in Q^* with area. Figure 10C shows
558 the distribution of k_{sn} values that meet these criteria. We found that k_{sn} was nearly twice

559 as high at Druids Run, with a median of 2.774 (IQR 2.163 – 3.284), as Baisman Run,
 560 with a median of 5.23 (IQR 4.747 – 7.017).

	k_{sn} (m)			Exponents (-)		Runoff (-)
	Median	LQ	UQ	m	n	Q_{max}^*
Druids Run	2.774	2.163	3.284	0.5	1	0.3
Baisman Run	5.230	4.747	7.017	0.5	1	0.3

Table 5. Channel steepness index k_{sn} , streampower exponents, and maximum runoff rate Q_{max}^* for Baisman Run and Druids Run.

561 We estimated the maximum dimensionless discharge Q_{max}^* at Baisman Run as the
 562 long-term average runoff ratio $\langle Q \rangle / \langle P \rangle = 0.3$ (Cosans, 2022). From our short timeseries
 563 at Druids Run, we calculated a runoff ratio of 0.57. Because k_{sn} depends on the prod-
 564 uct of K and Q_{max}^* (Equation 21) in our model, these data suggest that the factor of
 565 two difference in k_{sn} between our sites could be due to the difference in the hydrology,
 566 expressed in Q_{max}^* , rather than a difference in material and geomorphic properties, ex-
 567 pressed in K . While that would support our hypothesis, we will conservatively set $Q_{max}^* =$
 568 0.3 for Druids Run as a first estimate, matching Baisman Run.

569 With all components of Equation 22 estimated, we used the same Monte Carlo pro-
 570 cedure to calculate K and its uncertainty. Figure 10D shows that K is substantially higher
 571 at Druids Run than at Baisman Run when Q_{max}^* is set equal. The median at Druids Run
 572 is $1.34e-5 \text{ yr}^{-1}$ (IQR $8.24e-6 - 1.98e-5$), while at Baisman Run it is $6.49e-6 \text{ yr}^{-1}$ (IQR
 573 $3.83e-6 - 9.66e-6$). The full table of geomorphic parameters are shown in Table 6.

Name	Symbol	Units	Druids Run	Baisman Run
Uplift rate	U	m/yr	$1.143e-5$	$1.143e-5$
Hillslope diffusivity	D	m^2/yr	$8.611e-3$	$9.285e-3$
Streampower incision coefficient	K	1/yr	$1.334e-5$	$6.546e-6$
Contour length	v_0	m	30	30

Table 6. Geomorphic parameters needed to run DupuitLEM. We used the median value from the estimated parameter distributions for U , D , and K . The values for U and the characteristic contour length v_0 are identical at the two sites.

574 The difference in streampower incision coefficient between the two sites potentially
 575 confounds our interpretation of subsurface hydrologic controls on emergent hillslope length
 576 and hydrological function, assuming the difference is due to a contrast in material prop-
 577 erties rather than hydrology. Our estimated subsurface hydrological variables support
 578 our perceptual model of how the sites should be different if they have coevolved with their
 579 hydrology; lower transmissivity at Druids Run should lead to more surface runoff and
 580 channel incision, and greater extent of variably saturated areas than the high transmis-
 581 sivity conditions at Baisman Run. However, a higher streampower incision coefficient
 582 may indicate that runoff is more effective at detaching and transporting sediment out
 583 of the watershed at Druids Run, which could also lead to closer spacing of channels and
 584 shorter hillslopes (Perron et al., 2008).

585 To test whether subsurface hydrology is necessary and sufficient for explaining the
 586 difference in variable source areas and hillslope length at the two sites, we ran four sim-
 587 ulations, shown in Figure 11: two that represent our best estimates of hydrological and

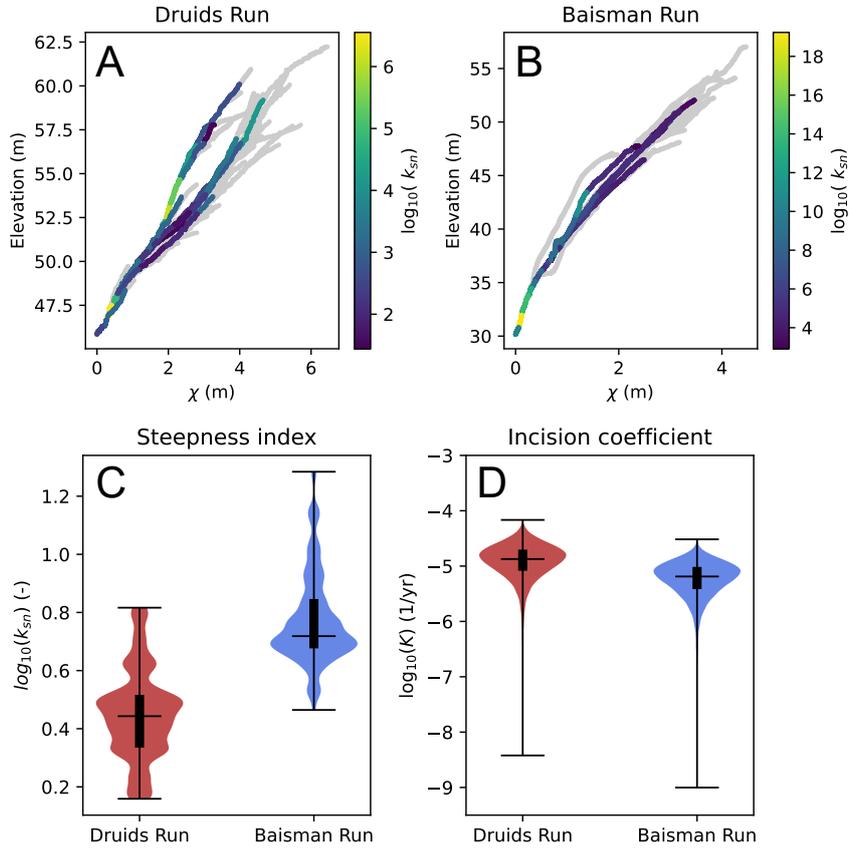


Figure 10. χ -elevation plots for Druids Run (A) and Baisman Run (B) for a concavity index $m/n = 0.5$. Channel segments are colored by their steepness index k_{sn} where the upslope area is greater than the 40th watershed area percentile, and are otherwise gray. (C) the distributions of k_{sn} for the segments colored in (A) and (B), showing generally higher channel steepness at Baisman Run than Druids Run. (D) distributions of the streampower incision coefficient K from Monte Carlo simulations. k_{sn} scales inversely with the erodibility, such that the streampower incision coefficient is lower at Baisman Run than Druids Run.

588 geomorphic parameters as described above (DR-DR, BR-BR), and two where we swapped
 589 the geomorphic parameters (DR-BR, BR-DR). Our best estimate cases helped discrim-
 590 inate how well DupuitLEM can capture landscape geomorphic and hydrologic dynam-
 591 ics at our sites. By comparing the best estimate simulations with simulations that have
 592 the same hydrological parameters but swapped geomorphic parameters, we determined
 593 whether geomorphic process rates alone explained the differences in morphology when
 594 the landscape coevolves with hydrology. Because we started with a randomized rough
 595 surface as an initial condition, we did not expect the simulation results to look exactly
 596 like Druids Run or Baisman Run. Instead, we compared them on the basis of aggregate
 597 properties including the hillslope length and relief, and saturation behavior.

		Hydrologic Variables (k_s, b, n_o)	
		Druids Run (DR)	Baisman Run (BR)
Geomorphic Variables (K, D)	DR	DR-DR	BR-DR
	BR	DR-BR	BR-BR

Figure 11. Four boxes indicating the four simulations we conducted. Colored boxes indicate the correctly matched hydrologic and geomorphic parameters, while white boxes indicate the ones in which the geomorphic variables are swapped. The listed hydrological and geomorphic variables are those that are varied, while all others are kept the same.

598 Lastly, we considered what happens when the differences in observed channel steep-
 599 ness were due to differences in runoff ratio (Q_{max}^*) rather than material properties (K).
 600 In our model formulation, determining the right value of Q_{max}^* should be an iterative
 601 process, in which the value of Q_{max}^* is estimated in order to determine erodibility, the
 602 model is run forward, the discharge and precipitation from the simulated landscape are
 603 used to recalculate Q_{max}^* , and then the streampower incision coefficient is adjusted ac-
 604 cordingly. This would be repeated until the estimated Q_{max}^* value matches the value pro-
 605 duced by the simulation. If there is a mismatch, the channel steepness of the modeled
 606 topography will be offset from that measured at the site. While we did not do a com-
 607 plete iterative solution, we did adjust Q_{max}^* and K according to the results of our first
 608 simulation.

609 3.3 Landscape evolution results

610 The landscape evolution model results showed the important effect of subsurface
 611 hydrology on the emergent landscapes, and revealed the complexity of interactions be-
 612 tween hydrologic and geomorphic processes. We first simulated topography for the four
 613 cases presented in Figure 11, and analyzed the hillslope properties and persistence of sat-
 614 urated areas using the same criteria as we used for the field sites. The only necessary
 615 difference was that we identified channel heads using a threshold on topographic curva-
 616 ture ($\nabla^2 z > 0.001$), because the DrEICH algorithm performed poorly on our model sim-
 617 ulations, which are much lower resolution than the lidar-derived DEMs. Because the trans-
 618 missivity is the primary difference in hydrological variables, we call the cases with hy-
 619 drology like Druids Run (DR-DR and DR-BR) the low transmissivity cases, and cases
 620 with hydrology like Baisman Run (BR-BR and BR-DR) the high transmissivity cases.

621 The most striking pattern in the hillshades shown in Figure 12A is that the low
 622 transmissivity cases were substantially more dissected than the high transmissivity cases.
 623 DR-DR and DR-BR have extensive fluvial dissection that extends onto hillslopes, which
 624 appears more extensive than we observed at Druids Run. However, the broad undissected
 625 hillslopes in BR-BR and BR-DR are similar to what we observed at Baisman Run. De-
 626 spite some visual similarities, Figure 12B–C shows that BR-BR and BR-DR cases tended
 627 to overpredict hillslope length and relief. Also, contrary to our expectations, in the low
 628 transmissivity cases where the geomorphic properties have been swapped (DR-DR ver-
 629 sus DR-BR), the difference in hillslope length and relief appeared to be comparable to
 630 the difference between Baisman Run and Druids Run (for a better view of length and
 631 relief at the field sites, see Figure 7). However, the presence of fluvial dissection broadly
 632 across these modeled topographies makes direct comparison with our field sites more dif-
 633 ficult. When the transmissivity is large, the channel network is very well defined, and
 634 we found less apparent effect of the difference in geomorphic parameters. While the 25th
 635 and 75th percentiles of hillslope length at BR-DR are smaller than those at BR-BR, their
 636 medians are approximately the same (Figure 12B).

637 Swapping geomorphic parameters had a relatively minor effect on hydrological func-
 638 tion. Figure 13A shows that simulations with swapped geomorphic parameters but the
 639 same hydrologic parameters have very similar saturated area patterns, whereas there is
 640 a substantial difference between simulations that have different hydrologic parameters.
 641 The low transmissivity cases have large variably saturated areas that extend onto hill-
 642 tops, as at Druids Run, though there are no hilltops that are classified as dry in the low
 643 transmissivity cases. They also show more persistent saturation in valley bottoms and
 644 zero-order basins than observed in Druids Run (Figure 13A–B). The saturated areas mod-
 645 eled in the high transmissivity cases look very similar to those observed at Baisman Run,
 646 where there is persistent saturation in valley bottoms and dry hilltops. The fractional
 647 saturated areas are similar to those observed at the sites as well (Figure 13B).

648 Next we examined the emergent runoff ratio and adjusted the fluvial parameters
 649 to account for the difference between the runoff ratio and the initial estimate of Q_{max}^* .
 650 The emergent runoff ratio for the high transmissivity cases were 0.33 and 0.32 for BR-
 651 BR and BR-DR respectively, which were very close to our initial estimate of 0.3, which
 652 was the observed runoff ratio at Baisman Run. The difference in geomorphic paramet-
 653 ers had little effect on emergent runoff ratio in these cases. In the low transmissivity
 654 cases, the runoff ratio was significantly higher than our initial estimate of 0.3. We found
 655 runoff ratios of 0.86 and 0.81 for DR-DR and DR-BR respectively. These values are again
 656 not highly sensitive to the difference in geomorphic parameters, but both are substan-
 657 tially higher than our initial estimate, and higher than our field estimate of 0.57 for Druid
 658 Run. However, this is consistent with our observation that DR-DR and DR-BR have much
 659 more extensive saturated areas than Druids Run. These higher runoff ratios suggest that
 660 we should increase estimated Q_{max}^* , and therefore decrease the estimated K at Druids
 661 Run. If we increase Q_{max}^* to 0.6, the corresponding K values is $6.68e-6 \text{ yr}^{-1}$, which is
 662 within 3% of the K value we estimated for Baisman Run. The geomorphic results of this
 663 increase are shown in Figure 14. The hydrologic effect of this increase is minimal, as shown
 664 in Figure S2.

665 Adjusting the streampower incision coefficient for differences in Q_{max}^* nearly elim-
 666 inates the difference in emergent morphology and hydrology between cases with swapped
 667 geomorphic parameters. The hydrological function of the landscapes is very similar when
 668 geomorphic parameters are swapped, which is expected given that there was little dif-
 669 ference in hydrological function between the original cases with swapped geomorphic pa-
 670 rameters. The emergent runoff ratio for DR-DR is now 0.78, which is slightly lower than
 671 we calculated previously. The emergent topography looks very similar when geomorphic
 672 parameters are swapped, and distributions of hillslope length and relief are nearly iden-
 673 tical (Figure 14). This suggests that the differences in the geomorphic parameters, and

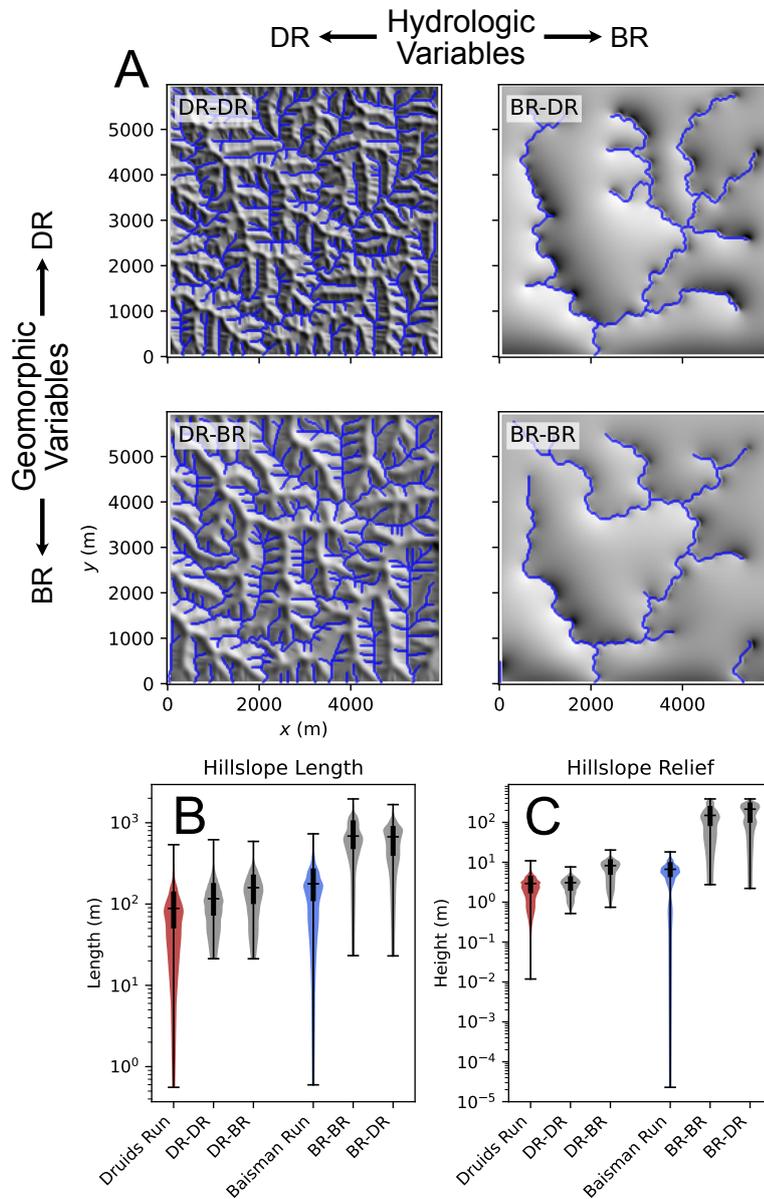


Figure 12. (A) Hillshades of model results in the same configuration as shown in Figure 11. Dissection is substantially higher in cases with Druids Run hydrological variables than Baisman Run hydrological variables. (B, C) Log-scaled violin plots of hillslope length and relief, comparing the field data (labelled “Druids Run” and “Baisman Run”) to the four modeled cases. Horizontal lines represent the maximum and minimum values, while the vertical bar represents the interquartile range.

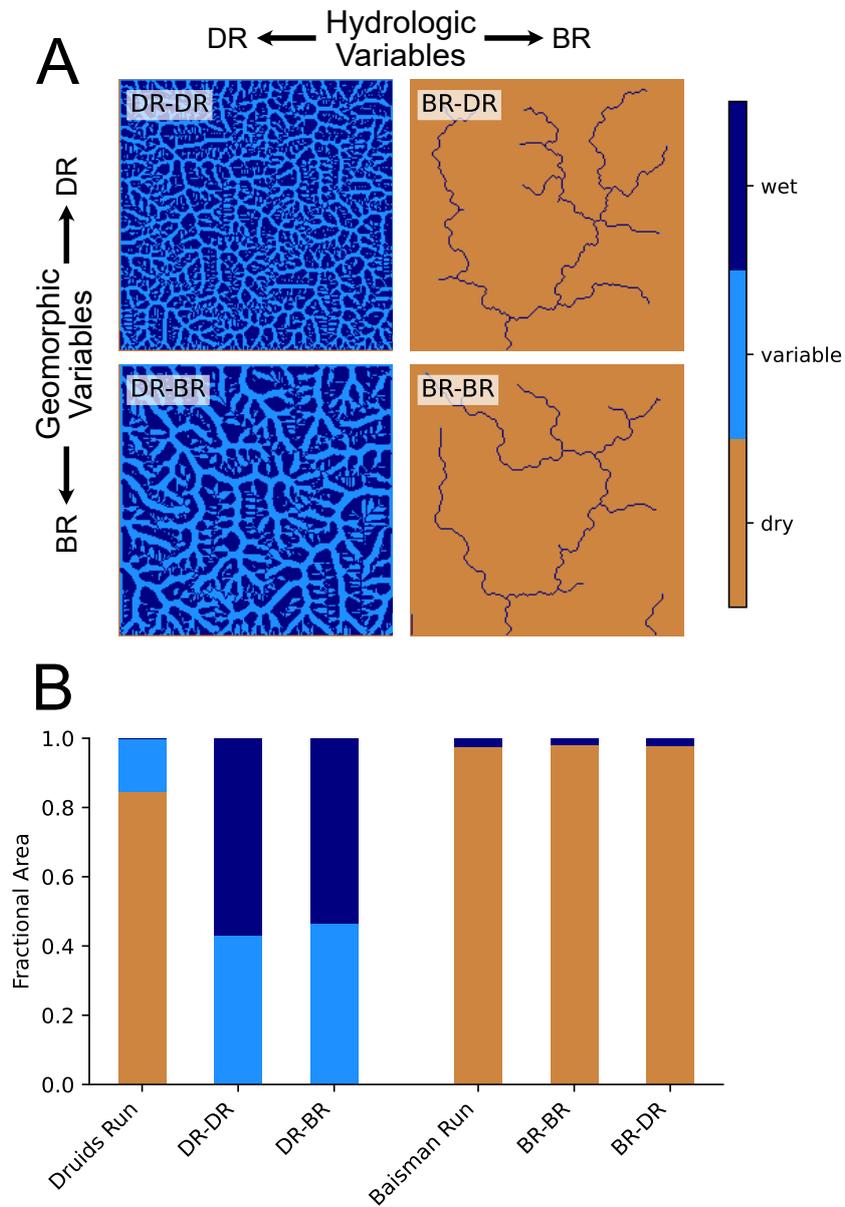


Figure 13. (A) Map view of saturated area classes for model results in the same configuration as shown in Figure 11 and Figure 12A. Saturated area behavior is not highly sensitive to swapping geomorphologic variables, while it is sensitive to swapping hydrological variables. (B) Fractional area that is classified as wet, variable, and saturated based on field data (labelled “Druids Run” and “Baisman Run”) and the four modeled cases. Cases that have the hydrological variables associated with Baisman Run appear similar to the field characteristics of Baisman Run. Cases that have the hydrological variables associated with Druids Run show more persistent saturation than the field characteristics of Druids Run.

674 in particular the intrinsic erodibility of the rock and regolith, are not responsible for the
 675 differences in emergent morphology. Instead, what we see in the difference in morphol-
 676 ogy between Druids Run and Baisman Run is more likely a combination of effects driven
 677 by the difference in their subsurface hydrology, as (1) the difference in transmissivity changes
 678 the extent of saturated areas and surface water on the landscape, which changes the pro-
 679 portion of the landscape that experiences fluvial erosion, and (2) higher runoff ratios in-
 680 crease the efficiency of water-driven sediment transport in areas where there is satura-
 681 tion, which further incises the landscape.

682 Our results also showed that there is more work to do to understand the controls
 683 on the geomorphic evolution of our sites. For instance, adjusting Q_{max}^* did not bring us
 684 closer to the true hillslope length and relief. Figure 15 shows how the true cases DR-DR
 685 and BR-BR compare to the hillslope length and relief of Druids Run and Baisman Run,
 686 respectively. The number in parentheses following the model label is the estimated value
 687 of Q_{max}^* . The values of hillslope length and relief from simulation DR-DR (0.6) were far-
 688 ther from the true values at Druids Run than those from simulation DR-DR (0.3). At
 689 the same time, we know that the channel steepness k_{sn} from the simulation DR-DR (0.3)
 690 will not match k_{sn} of Druids Run, because we overestimated the streampower incision
 691 coefficient K relative to the emergent value of Q_{max}^* . More work is needed to understand
 692 both the possible difference in other parameters (e.g., the denudation rate) and limita-
 693 tions of model structure for capturing our sites, but it is clear that the difference in the
 694 hydrology of the sites is an important component of their geomorphic evolution.

695 4 Discussion

696 4.1 The expression of subsurface hydrology in landscape evolution

697 It is well known that transmissivity affects the hydrological function of landscapes.
 698 All distributed hydrological models built on Darcy’s law will show a similar effect; the
 699 transmissivity, or more generally the depth-integrated hydraulic conductivity, affects the
 700 aquifer thickness and hydraulic gradient needed to convey a given water flux. This in
 701 turn determines how the water table will interact with the surface and produce overland
 702 flow (e.g., Beven & Kirkby, 1979; Li et al., 2014; Nippgen et al., 2015; Marçais et al., 2017).
 703 While there are limits to the Darcian approach for landscape scale runoff generation (e.g.,
 704 Uchida et al., 2005), it has proved useful for understanding and predicting runoff, sub-
 705 surface transport, and saturated areas.

706 Previous work toward understanding the role of transmissivity in topographic evo-
 707 lution (Luijendijk, 2022; Litwin et al., 2022; Litwin, Tucker, et al., 2023) is a logical ex-
 708 tension of the hydrological study of runoff generation, as sediment transport is an im-
 709 portant consequence of runoff generation. It has only recently received attention, in part
 710 because considering the long-term effects of this coevolution is computationally intens-
 711 sive, and in part because it relies on subsurface properties that are hard to estimate. Of-
 712 ten, landscape evolution modelers select the minimally-complex model needed to explain
 713 their observations. As a result, they have often excluded subsurface hydrology, despite
 714 the widespread importance of subsurface flow for runoff generation (Wu et al., 2021). How-
 715 ever, we have shown here that there are some cases where the subsurface hydrology is
 716 indispensable for understanding the evolution of landscapes. The importance of subsur-
 717 face runoff generation for a particular application of a landscape evolution model is de-
 718 pendent on the geologic and climatic setting, but also on the scale of interest. Studies
 719 focusing on watershed scales of 1-10s of kilometers may find that capturing subsurface
 720 flow is essential, while these details may be less important in the evolution of entire oro-
 721 gens, where the length of subsurface flow paths relevant to runoff generation is shorter
 722 than the scales of geomorphic interest.

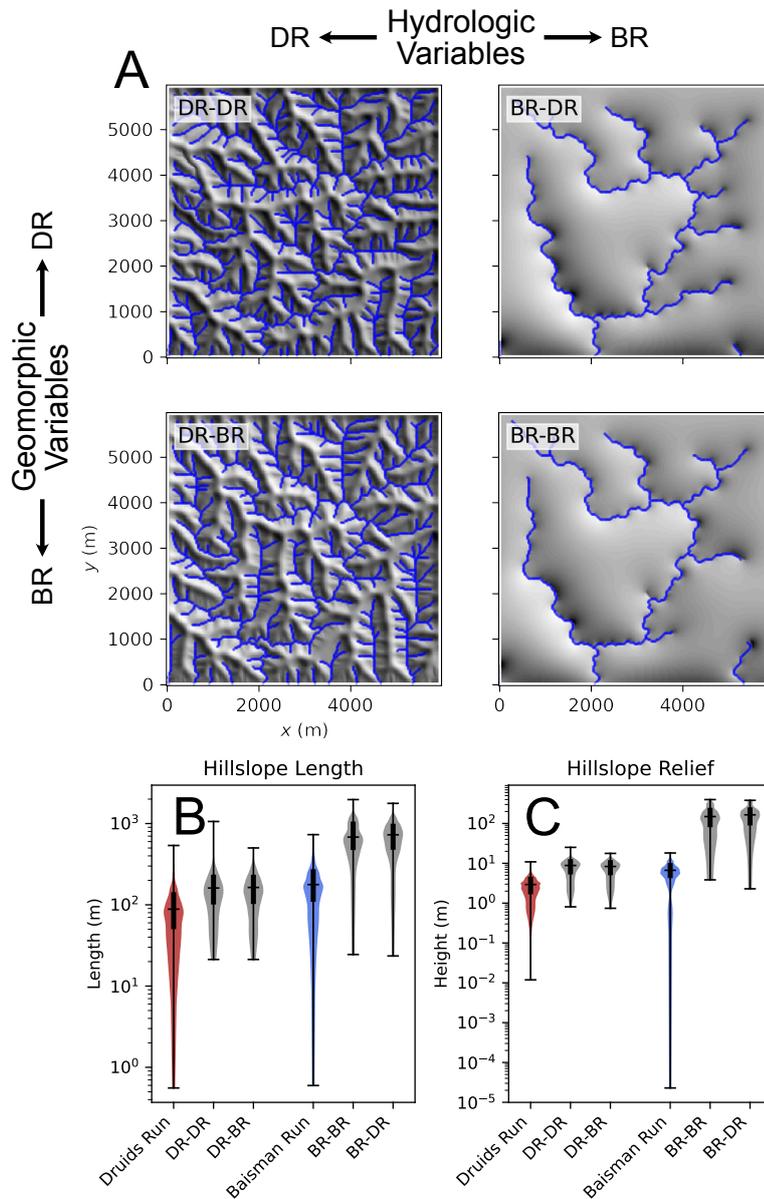


Figure 14. (A) Hillshades of model results in the same configuration as shown in Figure 11, only $Q_{max}^* = 0.6$ was used to determine the streampower incision coefficient for cases with Druids Run geomorphic variables. Visual comparison of results suggests that the difference in hydrology between the two sites is the primary control on emergent morphology. (B, C) Violin plots of hillslope length and relief, comparing the field data (labelled “Druids Run” and “Baisman Run”) to the four modeled cases. There is little difference between simulations with swapped geomorphic variables (comparing down columns), while there is still substantial sensitivity to swapped hydrological variables (compare across rows). All four modeled cases still have length and relief greater than those observed in the field.

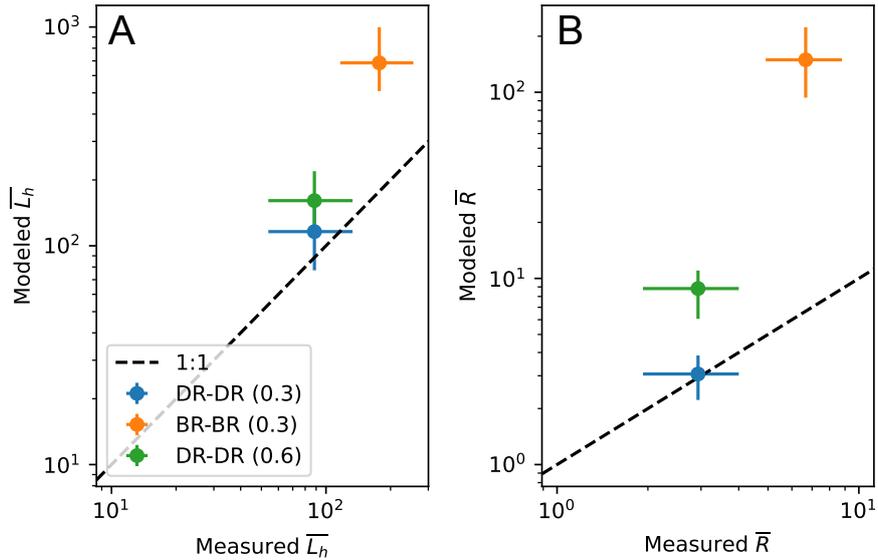


Figure 15. Modeled versus observed hillslope length (A) and hillslope relief (B). Results are only shown for the true cases (not the swapped parameter cases). The number following the simulation name is the value of Q_{max}^* . For Druids Run, simulations using both the original and updated estimates of Q_{max}^* are shown. The points are median values, and the error bars show the interquartile range.

723 We were able to show that hydrology was indispensable for understanding our sites,
 724 in part because we had hydrologic data for comparison, rather than just surface topog-
 725 raphy. Adding the hydrological dimension can help get the right answer for the right rea-
 726 sons in landscape evolution models, or discriminate when we have not gotten the right
 727 answer for the right reasons. This kind of approach could be useful beyond one-to-one
 728 site comparisons. For instance, we might be able to examine the effects of hydrological
 729 versus geomorphic processes on landscape evolution and hydrologic function across land-
 730 scapes with different lithologies, by using rainfall-runoff relationships or other hydrolog-
 731 ical indicators that are more widely available than saturation. Constraining subsurface
 732 properties is still challenging, but methods like the logistic regression approach we pre-
 733 sented here may be useful, especially as they are improved and refined.

734 4.2 Parameter estimation and limits of DupuitLEM

735 While our results provide evidence for a critical link between subsurface hydrolog-
 736 y and landscape evolution, there are clear discrepancies between the characteristics
 737 of Baisman Run and Druids Run that we observed and those we were able to model with
 738 DupuitLEM. Some of these discrepancies could be due to our choice of model param-
 739 eters, while others appear to be structural limitations of DupuitLEM.

740 Our results here and in prior studies (Litwin et al., 2022; Litwin, Tucker, et al., 2023)
 741 demonstrate that emergent topography and hydrology are highly sensitive to transmis-
 742 sivity, so the accuracy of the transmissivity estimate is likely a factor in model-data dis-
 743 crepancies. Our novel approach to estimate transmissivity relied on topographic index
 744 as a measure of hydrological similarity (O’Loughlin, 1986; Beven & Kirkby, 1979). How-
 745 ever, our results showed that topographic index and discharge, when combined in Equa-

tion 7, were only modestly good predictors of saturated area (Figure 8). Furthermore, topographic index is a resolution-dependent quantity (Zhang & Montgomery, 1994), which means that the resulting transmissivity that we calculate also depends on DEM resolution. While accounting for this effect is unlikely to change the relative magnitudes of transmissivity between the sites, it may change the estimated values. This was a problem with calibrated transmissivities in TOPMODEL as well (Beven, 1997), so some of the strategies that have been devised to reduce the scale dependence in that context (e.g., Saulnier et al., 1997) may be useful for improving our transmissivity estimates as well.

Our model results also showed that hillslope length and relief were too large in the simulated landscapes regardless of transmissivity. This could suggest that the relative magnitude of hillslope diffusivity to the fluvial erosion efficiency is too large (Perron et al., 2008; Theodoratos et al., 2018). Our modelled cases are generally able to reproduce observed hilltop curvature (Figure S3A), which suggests that the diffusivity is not the primary issue. Modelled channel steepness, however, is systematically larger than the channels from which the parameters were defined (Figure S3B). One likely issue that could explain this discrepancy arises from using K estimates from 1D channel profiles in a 2D model. Hillslopes in the 2D model contribute material to valleys that rivers must remove. This decreases their erosional efficiency compared to what is expected when estimating K from a 1D profile in which the river only needs to erode at a rate U (Equation 16). This topic requires further exploration than can be accommodated here, and will be covered in future work.

While there are limitations to our ability to estimate transmissivity and other process rates, we know that some key hydrological and geomorphic processes and features are missing from DupuitLEM. For example, DupuitLEM does not have a pathway for evaporation or transpiration of water once it has reached the saturated zone. Especially in cases where the water table is close to the surface, evaporation of saturated zone water is likely a significant control on hydrologic dynamics. Including it would decrease the proportion of the watershed that stays saturated during interstorm periods and decrease antecedent wetness when storms arrive.

Our simulations were also limited to cases where the subsurface thickness is uniform across the landscape. We know this may not generally be the case. In Baisman Run, deeply weathered zones under hillslopes delay the arrival of hillslope water to streams and support baseflow, while a relatively shallow subsurface in valley bottoms may increase the likelihood of overland flow in the channels (Cosans, 2022; St. Clair et al., 2015). This pattern could increase flow persistence and drainage dissection relative to a uniform subsurface. In contrast, very thin soils on hillslopes at Druids Run allow saturation and overland flow to occur frequently, while a more permeable valley bottom may increase the subsurface conveyance in valleys relative to the amount of water that remains after storms. Depending on how the riparian area is connected to the stream, it may also store more water that can be slowly released during interstorm periods. These patterns could increase or decrease saturated areas and drainage dissection, depending on the extent of the riparian aquifer and its stream connection. In addition to shaping subsurface structure, weathering can also result in significant chemical denudation. DupuitLEM, like many geomorphic models, assumes all denudation occurs by physical erosion. This limitation is discussed in the next section.

4.3 Chemical weathering and landscape evolution in the Eastern Piedmont

We were not the first to be interested in the contrast between landscapes on schist and serpentine bedrock in the Piedmont. Cleaves et al. (1974) also made this comparison, using Pond Branch (a subwatershed of Baisman Run), and a small watershed on the Soldiers Delight Ultramafite that is south of our site. Pond Branch had been stud-

797 ied previously (Cleaves et al., 1970), while this paper introduced the Soldiers Delight study
798 site. The focus of their study was to contrast the roles of physical versus chemical de-
799 nudation in the two terrains, and identify the hydrological signature of the deep sapro-
800 lite that is present on the schist. Cleaves et al. (1974) identified a hydrological signature
801 very similar to what we observed: the schist bedrock site had baseflow-dominated runoff,
802 which was highly persistent through droughts due to the large volume of storage in the
803 saprolite. In contrast, they observed that the site on the ultramafic bedrock generated
804 more quickflow, and had highly variable baseflow discharge, which they attributed to the
805 lack of saprolite.

806 In examining the morphologies of the two sites, they determined that there was no
807 strong evidence for differences in overall rates of denudation between the terrains. How-
808 ever, on the basis of a geochemical mass balance they determined that chemical weath-
809 ering was responsible for approximately 90% of denudation in Soldiers Delight at present,
810 while it was responsible for approximately 50% of denudation at Pond Branch. This would
811 suggest a significant difference in how we should interpret the resulting morphologies,
812 as we assumed that all denudation was due to physical erosion. Some recent work be-
813 gins to provide a framework for understanding morphologic effects of chemical denuda-
814 tion. Ben-Asher et al. (2019) introduced a modification of the hillslope mass balance that
815 includes chemical denudation in the form of a chemical depletion fraction (CDF). They
816 showed that curvature should be reduced as the ratio of chemical to total denudation
817 increases, assuming a constant hillslope diffusivity. Marcon (2019) applied this princi-
818 ple to several hillslopes on contrasting lithologies across the Piedmont, including sites
819 on schist and serpentine bedrock. They found decreasing hilltop curvature with increas-
820 ing CDF, where serpentine sites had the highest CDF and lowest hilltop curvatures. In-
821 terestingly, at our sites we found virtually no difference in hilltop curvature between litholo-
822 gies. If the total denudation rate at both sites is indeed very similar, but chemical de-
823 nudation is dramatically different, we are left with the conclusion that the identical cur-
824 vature is a coincidence that arises from higher hillslope diffusivity D at Druids Run than
825 Baisman Run. Dissolution could also have significant effects on river profiles, with con-
826 sequences for interpretations of channel steepness and streampower incision coefficient
827 K . Further research, including updated denudation estimates specific to our sites, would
828 be needed to draw further conclusions.

829 4.4 Toward surface–subsurface critical zone evolution

830 Here we sought to understand how subsurface properties condition runoff gener-
831 ation and, as a result, topographic evolution. Many other processes that build and struc-
832 ture the critical zone, which extends from the top of the canopy to the bottom of the cir-
833 culating groundwater, are closely coupled (Brantley et al., 2007; Troch et al., 2015). Be-
834 sides the hydrological link, other studies have shown that subsurface properties can in-
835 fluence morphology by limiting the size and effectiveness of sediment to do erosional work
836 (Callahan et al., 2019; Brocard et al., 2016). Others have investigated critical zone evo-
837 lution from fresh bedrock up to the surface, and have shown some aspects of how climate,
838 hydrology, and geomorphology condition the evolution of the subsurface (e.g., Rempe
839 & Dietrich, 2014; Harman & Cosans, 2019; Anderson et al., 2019). There is a lot of work
840 to be done to understand the feedbacks between surface and subsurface evolution, and
841 how they produce, across a range of climates and lithologies, relatively similar patterns
842 of drainage networks and soil-mantled hillslopes. At the same time, we have an increas-
843 ingly detailed picture of critical zone structure. High-resolution topographic data has
844 given us detailed insights into geomorphic processes acting at the surface (Sofia, 2020),
845 while near-surface geophysics has allowed us to peer into the subsurface and begin to test
846 models of subsurface evolution (Riebe et al., 2017; Parsekian et al., 2021). Our hope is
847 that future work will consider the importance of feedbacks between subsurface hydro-
848 logy and topography as we go forward in our understanding of critical zone structure and
849 evolution.

850 5 Conclusions

851 We framed this paper with two hypotheses about how the morphology and hydro-
 852 logical function of two landscapes should be different, informed by our understanding
 853 of how the subsurface affects coevolution of runoff and topography. We found that both
 854 the hydrological function and morphology aligned with our predictions. Druids Run, which
 855 has a thin permeable subsurface, had more extensive variably saturated areas, more vari-
 856 able effective area contributing runoff, and shorter hillslopes than Baisman Run, which
 857 has a deep permeable subsurface. An analysis of the available field data further showed
 858 that the transmissivity was substantially higher at Baisman Run than Druids Run. While
 859 these findings support our hypothesis that coevolution with subsurface hydrology is im-
 860 portant for emergent morphology and hydrological function, they did not in themselves
 861 provide a causal link. To test that link, we used a landscape evolution model with ground-
 862 water flow to show that the differences in geomorphic process rates (the hillslope diffu-
 863 sivity and streampower incision coefficient) were insufficient to explain the differences
 864 in morphology and hydrological function we observed. At the same time, we found dis-
 865 crepancies between the calibration-free model results and field data, which we discussed
 866 in the context of both parameter estimation challenges and model structure. The meth-
 867 ods we explored here could serve as the basis for future study to uncover the importance
 868 of subsurface hydrology for the evolution and hydrological function of landscapes.

869 6 Open Research

870 All original data, model output, and scripts needed to process data and generate
 871 figures are archived on Zenodo (Litwin & Harman, 2024). The Python package DupuitLEM
 872 v1.1 (Litwin, Barnhart, et al., 2023) contains the models and scripts used to generate
 873 and post-process the model output. Landlab v2.0 (Barnhart et al., 2020) is a core de-
 874 pendency of DupuitLEM.

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