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Key Points:
- We calculated the potential downslope travel distance of interflow in perched water bodies in 17 different hillslopes
- Downslope travel distances from the 17 hillslopes ranged from around 1 m to several hundred meters
- The analysis revealed that a continuously perched saturated zone with downslope flow does not imply continuous connectivity to the stream

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Interflow Is Not Binary: A Continuous Shallow Perched Layer Does Not Imply Continuous Connectivity

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Abstract Hillslopes exert critical controls on the quality and quantity of downstream waters. To understand and model dominant headwater catchment processes, we need to estimate the relative importance of different runoff generation processes. In this work we analyze published data from studies of 17 hillslopes from a range of landscapes to better understand the relative role of interflow, that is, shallow lateral subsurface flow moving over a layer impeding percolation, in streamflow generation. For each slope, we calculated downslope interflow travel distances, that is, the potential distance a water parcel travels downslope above an impeding layer until it percolates into the impeding layer. The downslope travel distances for the 17 hillslopes ranged from around 1 m to several hundred meters. The vector analysis of downslope travel distances revealed that all but three hillslopes had slope lengths that were much longer than downslope travel distances. For the remaining 14 cases we could show that most water perched above a shallow impeding layer percolates through the impeding layer before reaching the valley or the stream channel. Thus, interflow usually contributes directly to valley water or streamflow only from the lower portions of the hillslope in most landscapes. A critical finding of our analysis is that a continuously perched saturated zone with downslope flow does not imply continuous connectivity to the stream. Such a continuous connectivity is the exception rather than the rule in most landscapes. Future hillslope and headwater processes and modeling studies will need to account for this.

1. Introduction

Understanding and modeling the hydrologic and chemical transport through particular hillslopes and headwater basins requires estimation of the relative importance of various streamflow generation processes as affected by pedology, lithology, topography, climate, and floral and faunal influences on near surface hydraulic conductivity distributions (e.g., Atkinson, 1978, Figure 1; Bachmair & Weiler, 2011; Weiler et al., 2006). However, process identification and understanding is hampered because typical hillslope hydrometric monitoring systems quantify time series of hillslope states, such as piezometric or moisture levels, rather than flow directions, magnitudes, or processes (Bracken et al., 2013; Richards, 1994). Consequently, our ability to infer processes from such measurements depends partly on the types and spatial densities of sensors and partly on the spatial scale of the investigation. Process-based hillslope and headwater models may close such a gap (e.g., Glaser et al., 2016; Hopp & McDonnell, 2009), but they require detailed data sets about soil physics, hillslope structure, and validation data, yet such models may eventually suffer from parameter identification issues. Our goal here is to use data from published hillslope studies conducted in a variety of landscapes to better understand the relative role of interflow (i.e., shallow lateral subsurface flow moving over a layer that impedes further percolation) in hillslope flow processes and streamflow generation using a parsimonious approach. We address the particular inferential danger posed when examining piezometric observations of shallow perched water tables above a near surface layer that impedes percolation and assuming hillslope-valley connectivity of the entire perched water body.

The utility and accuracy of the concept of hydrological connectivity, defined as the transport of water "from one part of the landscape to the other" (Bracken & Croke, 2007) to link hillslope processes with catchment response (Martínez-Carreras et al., 2016; McGuire & McDonnell, 2010; Nippgen et al., 2015), depends on our ability to identify and quantify such connectivity. Several hillslope conceptualizations and resulting modeling systems assume that once shallow soil water perches above a slope-parallel impeding layer, it will continue to move downslope by this pathway (Nippgen et al., 2015; Smith et al., 2013; Troch et al., 2003, 2004). Recent analyses of perched interflow observations have implied, if not explicitly proposed, that in

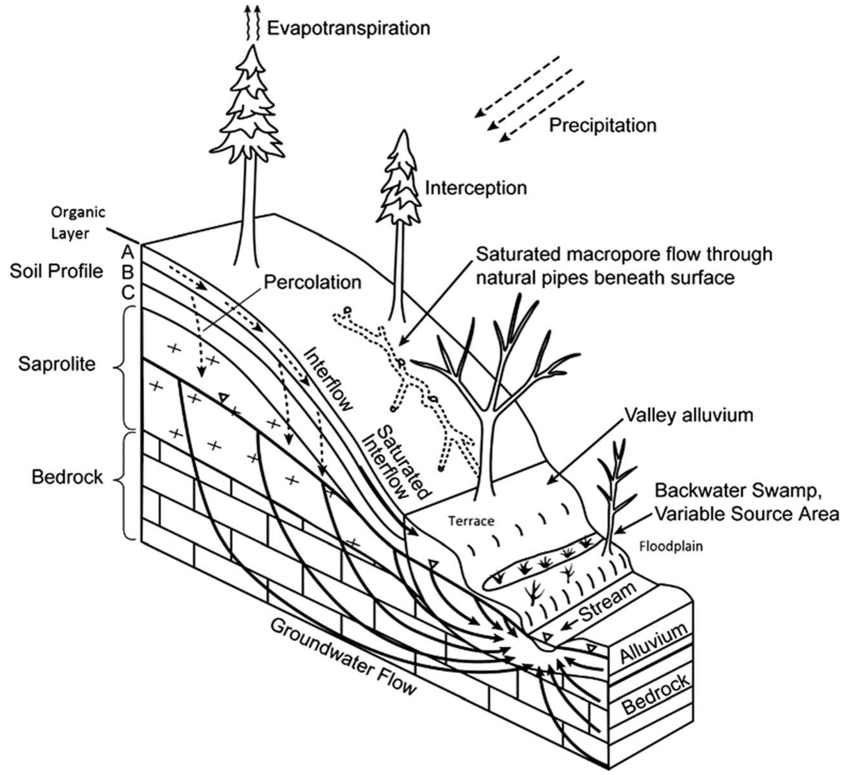

Figure 1. Generalization of hillslope flow paths in a typical humid forest with upland residual soils weathered from rock draining to an alluvial valley. In this hillslope example, the impeding layer is the argillic B horizon through which some percolation recharges the groundwater. Adapted from Atkinson (1978) and Jackson et al. (2014).

hillslopes all parts of a continuous perched aquifer are connected to streams or valleys (Emanuel et al., 2014; Jencso et al., 2009; Smith et al., 2013; Stieglitz et al., 2003; Vidon & Hill, 2004; Zimmer & McGlynn, 2017a, 2017b). Studying the joint dynamics of perched water above an argillic Bt horizon and a deeper water table in saprolite, Zimmer and McGlynn (2017a, 2017b) noted that runoff occurred with or without a measureable rise in the deeper water table and that shallow flow path activation was consistently coincident with runoff activation. They determined that perched interflow "created a hillslope-riparian-stream connection that is relatively common and independent of catchment storage state and deeper groundwater levels." Similarly, Jencso et al. (2009, 2010) assumed hydrologic connectivity of hillslopes and valleys wherever continuous perched shallow water tables were observed. Stieglitz et al. (2003) proposed that water table continuity eventually allows solute sources to connect to the stream. This concept, where shallow water table continuity between the uplands and the stream describes hydrological connectivity, was also used for a new hydrological modeling approach that was successfully applied in a steep, forested watershed for reproducing streamflow (Smith et al., 2013).

Such binary conceptualizations of interflow processes may be an artifact of the early experimental studies of interflow and the derived process understanding. Hewlett's inclined soil models, constructed to evaluate the potential role of interflow in maintaining base flow in mountain streams, featured a shallow soil layer within an impermeable concrete box (Hewlett, 1961; Hewlett & Hibbert, 1963). Later soil models also used impermeable boundaries (Anderson & Burt, 1977; Nieber & Walter, 1981). In this early work, downslope was the only direction soil water could move. Furthermore, many field studies of interflow have been conducted on steep slopes with shallow bedrock assumed, at least initially, to be nearly impermeable (cf. Weiler et al., 2006). This assumption is consistent with commonly observed transient saturation at the soil-bedrock interface during rainstorms and snowmelt in steep, humid terrain (Mosley, 1979), which leads to the initiation of lateral subsurface stormflow (Weyman, 1973; see also Tani, 1997) often through semiconnected meso- and macropores in the lower soil profile (Buttle & Turcotte, 1999; McDonnell, 1990; McGlynn et al., 2002). Although other mechanisms exist for subsurface stormflow initiation, like

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transmissivity feedbacks (Bishop, 1991; Detty & McGuire, 2010), lateral flow in the (unsaturated) upper soil profile (Tsuboyama et al., 1994), or fill and spill processes (Tromp-van Meerveld & McDonnell, 2006), the notion of impermeable bedrock controlling lateral subsurface stormflow is entrenched in the hillslope hydrological process literature (Gabrielli et al., 2012; Tromp-van Meerveld & McDonnell, 2006; Weiler et al., 2006). Recently, interflow studies have included hillslopes and headwaters with less steep slopes and have included measurements that revealed nonnegligible permeability of the underlying bedrock or impeding layer (Gabrielli et al., 2012, 2018; Graham, Woods, & McDonnell, 2010; Jackson et al., 2016; Klaus et al., 2015; McDaniel et al., 2008; Redding & Devito, 2008). Nevertheless, a binary conceptualization of interflow is often the process perception (e.g., Brantley et al., 2017) and is also coded into several modern numerical models of shallow lateral flow in hillslopes that consider the role of dynamic soil storage and evapotranspiration but assume no or minimal leakage through the impeding layer (e.g., Nippgen et al., 2015; Troch et al., 2003, 2004).

While this simplification is valid for artificial hillslopes and some real hillslopes, measurements of impeding layer conductivities or seepage rates—even for hillslopes with steep topography and hard crystalline rock near the surface—have challenged the validity of assuming that shallow bedrock or other impeding layers are impermeable. At the Panola Mountain research site, where competent granite lies a meter below the soil surface, 91% of applied irrigation water leaked out of the soil profile into the bedrock at steady state (Tromp-van Meerveld et al., 2007). A similar steady state irrigation experiment at the HJ Andrews experimental forest in Oregon found that around 27% of the applied water percolated through the andesite and coarse breccias underlying the soil and saprolite (Graham, Van Verseveld, et al., 2010). Graham, Woods, and McDonnell (2010) and Gabrielli et al. (2012, 2018) measured seepage rates and conductivities of the impeding Old Man Gravels at the MaiMai hillslopes and found potential for substantial percolation into the formation, although their estimates of impeding conductivities ranged over three orders of magnitude. Most of the rainfall was on a steep Japanese catchment overlying weathered granite percolated into the granite and recharged the deeper groundwater body (Katsuyama et al., 2005). Parlange et al. (1989) found that a low conductivity fragipan contained fracture features through which substantial amounts of perched water percolated. Based on long-term observations and modeling at an experimental hillslope in Czech Republic, Dusek and Vogel (2016) concluded that approximately 17% of precipitation during the growing season percolated into in the underlying granite bedrock. These findings are critical not only for the hillslope water balance but also for the travel distance, or contributing length, of interflow. Jackson et al. (2014) demonstrated through the addition of downslope and normal flow vectors that the downslope travel distance of perched interflow was proportional to the ratio of these vectors (Figure 2). In other words, the greater the leakage rate through the impeding layer, the shorter the contributing hillslope distance to the riparian aquifer. They suggested that in most landscapes interflow only reaches the valley from the lower portions of hillslopes. Similarly, Hrncir et al. (2010) found that their hillslope trench only drained 25–30 m of a hillslope (14% slope) above the trench, while the topographic watershed divide was located 130 m above their trench.

In this work, we will use Darcian principles (see Jackson et al., 2014) and published field data from an array of studies to show that the binary conceptualization of interflow is incorrect in most landscapes. We will show that it is far more common that most water perched over shallow impeding layers percolates through the impeding layer long before reaching the valley floor. Consequently, a continuously perched saturated lens over a shallow impeding layer does not imply continuous connection of shallow lateral flow to the stream valley.

2. Estimation of Downslope Travel Distances From Near-Surface and Impeding Layer Conductivities Reported From Studies Around the World

Per Jackson et al. (2014), the downslope travel distance (DTD) of perched interflow is estimated as the product of the ratio of the hydraulic conductivities of the upper layer (K_{upper}) and the impeding layer ($K_{impeding}$), the slope of the hillslope relative (I_h) to the normal hydraulic gradient (I_n), and the thickness (h) of the perched saturated (Figure 2):

$$\mathrm{DTD} = \frac{\mathrm{K_{upper}}}{\mathrm{K_{impeding}}} \times \frac{I_h}{I_h} \times h \tag{1}$$

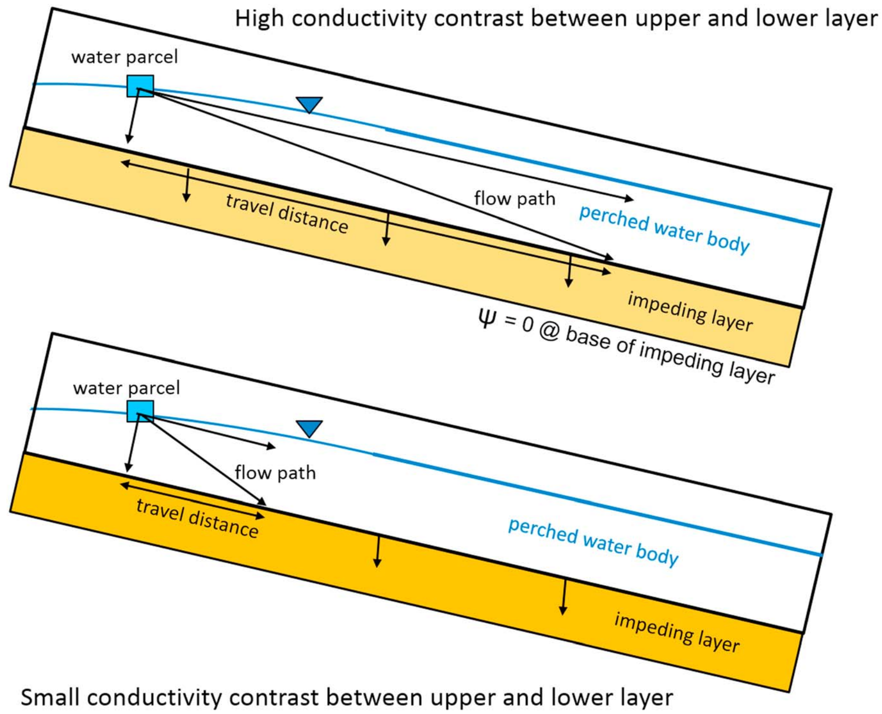

Figure 2. Vector diagram of flow components indicating the movement of a water parcel in a hillslope above an impeding layer not connected to an underlaying groundwater body. A slope-parallel flow vector and a normal percolation vector add up to a downslope vector of travel distance that depends on the slope, the conductivity contrast (two scenarios are pictured), and the perched water body depth. We assume isotropic hydraulic conductivities. Figure developed based on Jackson et al. (2014).

While layering of hydraulic conductivity can lead to capillary barriers and hydraulic barriers (Li et al., 2014) that interrupt percolation, this study evaluates only settings with a hydraulic barrier induced by a reduced hydraulic conductivities in the lower soil layer.

To calculate interflow travel distances and eventually upslope connectivity estimates of the hillslope properties (equation (1)) are required. Therefore, we reviewed the hillslope literature for papers that (i) reported perched interflow so that we could assume the pressure at the bottom of the impeding layer was zero and (ii) that also reported either a central tendency or a relatively narrow range of conductivities for the upper soil and the impeding layer and either an average or a range of slopes for the hillslope. When K ranges were provided, we calculated corresponding ranges of travel distances (as specified in Table 1). In some cases, conductivities were reported for more than one layer above the impeding layer, in which case we used a depth-weighted average conductivity above the impeding layer where possible. Some studies reported the relevant information for different segments of the hillslope, in which case we calculated travel distances for each segment. The studies used many different methods for estimating conductivities, and we provided notes on data sources and assumptions in Table 1. For calculating DTDs from these data, we assumed a normal hydraulic gradient of one through the impeding layer, and we used the maximum soil depth as the perching depth. Both of these assumptions produce high-biased estimates of the DTD.

We were able to find papers from 17 different field sites that met the perching requirement and provided the required information on slopes and hydraulic conductivities (Table 1). These studies span the globe (although the majority is from the USA) and include a variety of topographic, pedologic, lithologic, vegetative, and climatic conditions. We did not use studies where the interflow producing layer was connected to the water table because in such cases we cannot estimate a normal gradient. The perching requirement eliminated studies from the Canadian Shield, for example. We were also unable to use studies that reported large ranges of conductivities for each layer, but no central tendency. Hydraulic conductivities of surface soils and near-surface impeding layers can range over several orders of magnitude (e.g., Jackson et al., 2016), and therefore, ranges of hydraulic conductivities without central tendencies could not be used to estimate DTD.

Table 1

Perched Interflow Travel Distances Inferred From Published Values of Slopes, Upper (K_{upper}) and Impeding Layer ($K_{impeding}$) Conductivities, and Surface Soil Depths in a Variety of Settings

Reference	Interflow setting	K_{upper} (mm/hr)	$K_{impeding}$ (mm/hr)	K_{ratio}	Slope (fraction) and slope length (m)	Interflow travel distances
Arnett (1974)	Approximately 0.15 m deep A horizon with high organic matter over B horizon in the moors of Yorkshire, England	9–200	Not reported Permeabilities measured seasonally, max. ratio of permeabilities used for *K* ratio	Max. of 2.0 Very low *K* contrast determined from differences in permeabilities measured in falling head tests.	0.375 240 m	0.5 m
Betson et al. (1968)	Flow moves through an A horizon (depth of 0.3–0.6) over a denser AB horizon in agricultural fields in western NC, USA.	25–76 Range, no central tendency reported	8 Only one value reported	3.1–9.5 Low *K* contrast	0.4 Slope length not reported	0.8–2.3 m
Detty and McGuire (2010)	Shallow (0.7 m) Spodosols and Inceptisols over a glacial till C horizon on steep, humid, temperate forested slopes in White Mountains of New Hampshire. (WS1 in Hubbard Brook Experimental Forest)	13 (weighted average of medians of topsoil and B horizon conductivities measured with a constant head permeameter ~100 (high end of range)	6.3 (median value of C horizon conductivity measured with a constant head permeameter)	~2 using central tendencies, ~16 when perching reaches near-surface soils.[a] Low *K* contrast in general.[a]	0.30–0.35 50–300 m	Only about 1 m for shallow perching but may extend several meters when transmissivity threshold is met.[a]
Dusek and Vogel (2016), Hrncir et al. (2010), and Sanda et al. (2014)	Shallow (0.75 m) Cryptopodzol over weathered granite bedrock, covered by grass and spruce, Uhlirska, Czech Republic	34.3 (as depth-weighted average value of several layers)	0.17 (one value reported for each layer, determined from several cores)	202	0.14 130 m	19.8 m
Gabrielli et al. (2018, 2012), Graham, Woods, and McDonnell (2010), and McDonnell (1990)	Wet mixed evergreen forest silty loam soils overlying a poorly-permeable conglomerate known as the Old Man Gravel at 0.25–1.3 m depth (ave. 0.6) at the MaiMai catchment, New Zealand	5–300	0.20 (average of 12 hillslope slug tests) 2.4 (average of 11 hollow slug tests)	2.1–1500 Large ranges of topsoil and Bedrock K	0.67 average on hillslope. Range 0.27–2.14 Not reported for hollows. 100 m or less	10–603 m on hillslope. Variability due to large range of topsoil K. Uncalculated variability due to high slope variation Not calculated for low gradient hollows
Haught and Meerveld (2011) and Haught (2010)	Wet mixed, evergreen forest with sandy loams with large amount of woody debris in the soil underlain by cretaceous quartz diorite and granodiorite, max. soil depth of 1.1 m as point of refusal in piezometer installation), Malcom Knapp Research Forest	36 (at the soil bedrock interface)	0.036 given by the authors as summary from three bedrock infiltration experiments; a range over 3 orders of magnitude was reported)	1,000 Variability of bedrock K can lead to a high range of ratios (100–10,000)	0.53 Slope length not reported	583 m, ranges of bedrock conductivity can further increase travel distance

Table 1 *(continued)*

Reference	Interflow setting	K_{upper} (mm/hr)	$K_{impeding}$ (mm/hr)	K_{ratio}	Slope (fraction) and slope length (m)	Interflow travel distances
Jackson et al. (2014, 2016)	Perching occurs in loamy sands over an argillic Bt horizon at 0.2–2.0 m depth (ave. ~1) underlain by deltaic sands in humid, temperate, forested Ultisols of the Sandhills of the SE United States on low gradient hills.	460 Conductivities determined from plot-scale irrigation experiment	2.5 Conductivities determined from plot-scale irrigation experiment	184	0.02 to 0.20, typically around 0.08 200–300 m	Maximum values of approx. 36 m, 3.7 m minimum, 14.7 m typical, most less than 20 m
Jamison and Peters (1967)	0.28 m of A horizon soils over a claypan Bt horizon in pastures in Missouri, USA	497 (A horizon, below litter layer)	0.106 (B horizon) All K values determined from double-tube field measurements, reps not reported	4700	0.03 100 m	39.5 m
Katsuyama et al. (2005) and Ohte et al. (1989)	1.5 m of humid, temperate, forest soils over permeable weathered granite on moderate to steep slopes in the Matsuzawa Experimental catchment, Japan	1,692	21 (min. of 4 measurements)	81[b]	0.27–0.47 30–90 m	33–57 m
Knapp (1974)	1.77 m of solum over bedrock on grassland slopes of the Plynlimon watershed in central Wales	Not reported, but horiz/vert anisotropy ratio reported	Not reported	10 Horiz/vert anisotropy due to layering reported; this anisotropy ratio used as K ratio	0.23 Slope length not reported	4.1 m
Knoepp et al. (2012)	Interflow moves through high porosity, high organic matter content humid temperature forest soils overlying a denser saprolite at 0.5-m depth at the Coweeta Hydrologic Laboratory, southern Appalachian mountains.	234 (hill) 365 (vall) *K* values measured in-situ with a compact constant head permeameter, # reps not reported.	82 (hill) 87 (vall)	2.9 4.2 Very low *K* contrast	0.3 200–300 m	About 1 m or less
McDaniel et al. (2008)	Perching occurs in shallow (0.7 m) silt-loams overlying a Btxb fragipan in semiarid grasslands of the rolling Palouse Hills of eastern Washington.	8.13 (weighted average of Ap and Bw horizon conductivities measured with Guelph permeameter)	0.38 Permeameter measurements repeated at least nine times in each horizon.	21.4 Low *K* contrast	0.16 80–100 m	2.4 m
McGuire and McDonnell (2010), Harr (1977), and Ranken (1974)	Interflow moves through 1.3 m of humid, temperate, forested clay loam soils overlying 1–7 m of saprolite over andesitic bedrock on steep mountainous slopes of western Oregon.	1,675 (weighted average of first four horizons)	160 Conductivities determined in lab from over a thousand intact cores from the hillslope	10.5 Low *K* contrast	0.57–1.1 125 m	8–15 m

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Table 1 *(continued)*

Reference	Interflow setting	K_{upper} (mm/hr)	$K_{impeding}$ (mm/hr)	K_{ratio}	Slope (fraction) and slope length (m)	Interflow travel distances
	(HJ Andrews Experimental Forest).					
Newman et al. (1998) and Wilcox et al. (1997)	Desert southwest in New Mexico at Los Alamos. Flow moves through A and Bw horizons over Bt at 0.43-m depth. High conductivity A horizon is 6 cm thick, so long interflow distances only occur when soil profile is nearly saturated.	0.021–2.7 (A horizon, range, no central tendency reported)	$9*10^{-4}$ (Bt horiz, conductivities determined from lab measurements of large intact cores)	23–3,000	0.06 200–300 m	0.6–77 m Suggests transmissivity feedback similar to observations in Detty and McGuire (2010)
Tromp-van Meerveld et al. (2007) and Burns et al. (2003)	Humid, temperate, forested loamy soils averaging 0.63-m depth overlying 2–3 m of saprolite from weathered granite on moderate slopes of the Panola Mountain pluton, Georgia, USA	644 Conductivities determined from irrigation experiments	5.8	111	0.23 125 m	16 m
Whipkey (1965)	Interflow moves through sandy loam topsoil and a loam transition layer over a clay loam argillic horizon at 1.2-m depth in a mixed oak forest in the Allegheny Plateau in east-central Ohio.	286 Conductivity of the saturated portion of the sandy loam topsoils	2 Conductivities determined from irrigation experiments	143	0.28 100–300 m[c]	48 m
Zimmer and McGlynn (2017a, 2017b)	Perching occurs over an argillic Bt horizon at 0.6–1.05 m depth underlain by saprolite in humid, temperate, forested Ultisols of the Piedmont of the SE United States on gently rolling topography, Duke Forest Research Watershed.	17, weighted average of 4.4 and 29.3 in AB and A horizons in upper slope 26, weighted average of 4.2 and 47.5 in AB and A horizons i n lower slope	0.2 Upper slope 0.6 Lower slope	85 43	0.07 ~100 m	6.2 m Upper 3.2 m Lower Data also suggest transmissivity feedback similar to Detty and McGuire (2010).

Note. Travel distances are the product of the K_{ratio}, the slope, and the soil thickness above the impeding layer (Jackson et al., 2014). Travel distances in bold are longer than 50% of hillslope length. [a]Detty and McGuire determined that the effective upper conductivities increased as perched water table depths increased and water moved into highly transmissive upper soils. [b]Potentially high biased as only the minimum bedrock conductivity was reported. Their water balance findings suggest greater rates of percolation into bedrock. [c]Hillslope lengths not reported in the paper, representative lengths estimated from maps of the study area.

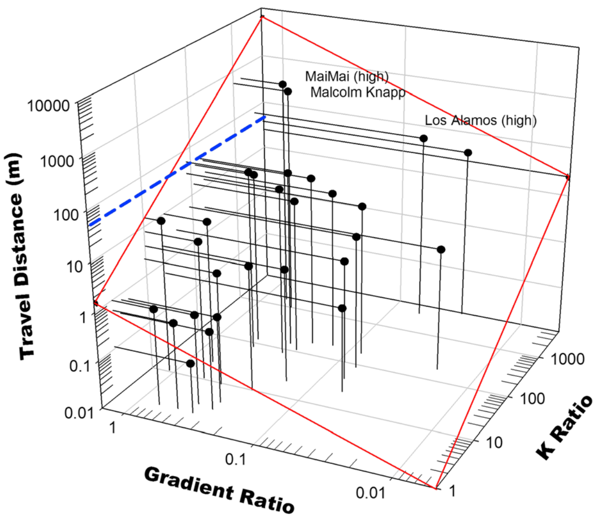

Figure 3. Downslope travel distances estimated for the 17 hillslopes in relationship to the variable space defined by reasonable ranges of the *K* ratio (from 1 to 5,000) and the gradient ratio (from.005 to 1.5). Travel distances for the "corners" (in red) of this variable space were calculated with a soil depth of 1 m. The dashed blue line marks a travel distance of 50 m. The data points are based on the interflow travel distances (Table 1). If ranges were presented in Table 1, the minimum and maximum values are plotted here.

3. Calculated Downslope Travel Distances in Different Landscapes and Their Relationship to Slope Distances

Downslope travel distances from the 17 study sites ranged from around 1 m to several hundred meters (Table 1 and Figure 3). For 13 of the 17 hillslope environments, the calculated maximum DTDs were shorter than 50 m; for 11, the calculated travel distances were less than 25 m; and for 9, the travel distances were less than 16 m. Except for three sites (the long interflow travel distance estimates for the Mai Mai, Matsuzawa catchment, and the Malcolm Knapp Research Forest), slope lengths were much longer than DTDs. For 14 of the 17 studied hillslope environments the calculated DTDs were less than 50% of slope length and 11 had DTDs less than 30% of slope length. In 14 of 17 cases, vector analysis of downslope and normal fluxes revealed that most water perched above a shallow impeding layer percolates through the impeding layer before reaching the valley. This shows that interflow contributes directly to the valley only from the lower portions of the hillslopes in most landscapes. These data, generated from studies around the world where interflow was expected to be an important runoff process, reveal that a continuously perched saturated zone with downslope flow does not imply continuous connectivity to the valley.

Plotting the DTDs from the reviewed studies against the range of travel distances calculated across reasonable values of the *K* ratio (from 1 to 5,000) and the gradient ratio (from 0.05 to 1.5) reveals that these 17 hillslopes actually span a substantial part of this variable space (Figure 3). *K* ratios in the study hillslopes varied from 2 to 4,700, and downslope gradients varied from 0.02 to 1.1. For good reason, hillslope researchers generally do not install interflow studies on flat lands. Six of these studies examined slopes with *K* ratios of around 10 or less. Such low *K* ratios are incapable of moving interflow very far, even with slopes approaching 100%. The depth of the high conductivity surface layers at these 17 hillslopes ranged from 0.15 to 1.77 m, with a median value of 0.7 m. Studies of root distributions from around the world have shown that almost all roots occur in the first 2 m (Schenk & Jackson, 2002). Thus, the potential thickness of the perched saturated zone is limited. Although these hillslopes were selected for the express purpose of studying interflow, only the Mai Mai and the Malcolm Knapp Research Forest hillslopes had a *K* ratio and gradient ratio combination sufficient for interflow to move more than 100-m downslope. In other words, hillslope hydrologists' intuition about the importance of interflow in hillslope environments appears to be high-biased. Interflow can move subsurface water a long way downslope in some hillslopes, but the DTD equation and the data from published hillslope studies indicate that such hillslopes are rare. Travel distances for these "interflow-producing" hillslopes clustered between ~10 and ~50 m substantial distances, but only fractions of slope lengths.

The inference from these estimated DTDs seems to match inference from hillslope water balances. The 19.8-m travel distance estimated for the Czech hillslope studied, for example, by Dusek and Vogel (2016) closely matched the 25-m contributing length estimated from their experimental work. Similarly, Jackson et al. (2016) also found that the contributing length estimated from the water balance for a plot-scale irrigation experiment also closely matched the estimated DTD within the R watershed at the Savannah River Site. Irrigating a hillslope underlain by a fractured and leaky fragipan, Parlange et al. (1989) found that the effective contributing area was much smaller than the irrigated area.

4. Limitations of Downslope Travel Distance Analysis

Interflow travel distance calculations are a simple and powerful tool for assessing the relative importance of interflow in a catchment and for differentiating the landscape with respect to dominant flow processes, but the calculation is limited to perched interflow environments above a hydraulic barrier. The advantage lies in the parsimonious approach that can classify hillslopes and landscape based on the DTD. This allows a fast and robust comparison of potential runoff generation mechanisms and hydrological connectivity in different watersheds, and it may assist in identifying "organizing principles that might underlie the heterogeneity and complexity" (McDonnell et al., 2007). In addition, one could evaluate the depth and duration of perching dynamics relative to duration and intensity characteristics of the local rainfall events and relative to downslope travel times to estimate the likelihood of interflow transport to the valley (e.g., Detty & McGuire, 2010; Jackson et al., 2014).

The calculations cannot be applied where interflow above an impeding layer is connected to the groundwater table without some measurement or assumption of the normal hydraulic gradient. For example, the method could not be applied to low relief glacial catchments studied by Vidon and Hill (2004) because data were not available to determine if the shallow water table was perched. We were surprised (and somewhat disappointed) that we could find studies of only 17 hillslopes that included all the data (topsoil conductivities, impeding layer conductivities, and evidence of perching) needed to estimate DTDs. Nevertheless, data from these 17 hillslopes, along with other studies that also measured conductivities in shallow bedrock between 7.2 and 54 mm/hr (Graham et al., 1997; Johnson-Maynard et al., 1994; Megahan & Clayton, 1986; Uchida et al., 2003), indicate that leaky impeding layers, whether they are weathered crystalline bedrock, argillic horizons, or till layers, are the norm, not the exception. Yet measuring the permeability of these layers remains a challenge.

Furthermore, the heterogeneity and anisotropy of soil hydraulic conductivities and differences in measurement techniques both complicate estimation of DTDs. Some of this heterogeneity is structured, as where there are multiple high conductivity soil layers at the surface (e.g., Dusek & Vogel, 2016). Heterogeneity in the hydraulic conductivity of the impeding layer can also modify DTD on small scales; for example, Du et al. (2016) found particularly locations with high conductivities in an otherwise low conductive argillic horizon that could lead to increased percolation in these spots; the same can apply to fractures in the impeding layer (Parlange et al., 1989). Conversely, flow paths with higher conductivity ratios could create tendrils of longer contributing zones extending farther upslope. Anisotropy between vertical and lateral hydraulic conductivities (Bouma & Dekker, 1981) may also influence the DTD analysis. Commonly, the observed lateral hydraulic conductivities are much greater than their vertical counterparts (Bathke & Cassel, 1991; Beckwith et al., 2003). This was not differentiated in the data sets used for these studies, but this effect may potentially increase the DTDs. Yet one can account for this process as long the necessary measurements are available for the investigated soils and hillslopes. Heterogeneities within the layering of soil horizons or subsurface hillslope structures introduce further complexity in the DTD analysis. Here we focused (also due to the data availability) on a permeable soil above one impeding layer. Obviously, one may find more complex systems, and the conductivities below the impeding layer can affect the analysis.

Tracer studies and other techniques often reveal preferential flow paths with conductivities much larger than the matrix conductivity (Anderson et al., 2009a, 2009b; Graham, Van Verseveld, et al., 2010; Jackson et al., 2016; Klaus et al., 2013; Retter et al., 2006). In such cases, we might conceptualize two characteristic interflow travel distances, one for matrix flow and one for preferential flow. This duality is a simplification, as there is a distribution of conductivities within any layer of the hillslope (Detty & McGuire, 2010; Jackson et al., 2016). Flow complexities will lead to a more complex pattern of DTDs potentially connecting some parts of the hillslope via lateral preferential flow to the valley bottom (as shown by Wilson et al., 2017), while nonpreferential interflow has much shorter DTDs. Within most hillslopes the DTD and thus the hydrological connectivity were far less than the hillslope length. Undetermined yet is the question of whether upslope infiltration and percolation could still play a role in interflow processes via pressure propagation (Rasmussen et al., 2000; Torres et al., 1998).

5. Conclusions

Here we used published hillslope data from around the world to show how the parsimonious DTD, which is the product of the conductivity ratio of the impeding layer and the soil above, the slope, and the thickness of the perched water above the impeding layer (Jackson et al., 2014), can be used to examine the relative importance of interflow in different landscapes. The concept of DTD is based on the fact that any water parcel on the top of a perched water table in a hillslope will eventually infiltrate through the underlying impeding layer given enough time when the conductivity of the impeding layer is not zero. Thus, the distance a water parcel can travel downslope in a perched water body is limited. This plays a critical role in assessing hydrological connectivity within a landscape and the associated runoff generation and solute transport processes. Here we show based on hillslope data from 17 published hillslope studies that calculated DTDs ranged from about one to several hundred meters, with the majority of hillslopes having DTDs less than 50 m. Even more, within the available data sets most landscapes had travel distances that are less than 30% of the hillslope length, which shows that most shallow perched water percolates through the impeding layer before contributing to valley water or streamflow via interflow. In other words, a continuously perched downgradient saturated zone within a hillslope does not imply that hydrological connectivity between the upslope and the valley exists. The extent of hydrological connectivity via interflow is limited by the potential DTD. In the majority of landscapes in this study only the lower portions of hillslopes could contribute to streamflow via interflow. Further work should explore the role of preferential flow and the distributions of hydraulic conductivities within soils on DTDs and should explore the travel distance concept in other landscapes. Future collaborative efforts in critical zone science will be necessary to understand the controls of complex hillslope structures on interflow generation and travel distances. Future work should also reassess the importance of hillslopes on catchment scale runoff generation within the framework of DTDs.

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