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REVIEW

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JAMES BUTTLE REVIEW: Interflow, subsurface stormflow and throughflow: A synthesis of field work and modelling

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Abstract

Interflow, throughflow and subsurface stormflow are interchangeable terms that refer to the lateral subsurface flow above a restricting layer of lower hydraulic conductivity that occurs during and following storm events. Interflow (used here) is a more dominant process in steeper catchments with high infiltration capacity soils overlying a more impermeable soil or geologic layer. Interflow as a runoff process was first recognised in the early 1900s, yet hydrologists still struggle to predict its occurrence, persistence, importance, interaction with other streamflow generation processes, and potential to connect to valleys and streams during and following storms. We review the history of interflow research and address some of the challenges in understanding its role in runoff production. We argue that characterising the controls on interflow initiation and occurrence relies on detailed field observations of subsurface properties, which exist only in limited experimental settings. This data shortcoming contributes to our inability to predict interflow or determine its contribution to streamflow more broadly. There remain many opportunities to advance our understanding of interflow that include both modelling and experimental or observational approaches in hydrology.

KEYWORDS

hillslope hydrology, interflow, lateral flow, runoff, streamflow generation, subsurface stormflow, throughflow

1 | INTRODUCTION

Interflow is a term used to represent lateral subsurface flow moving above a lower hydraulic conductivity soil, rock, or material of lithologic origin or pedogenic creation, which we will call the restrictive layer (Figure 1). Consequently, interflow becomes a more important hillslope flow pathway when slopes are steeper and there is high contrast between the hydraulic conductivities (K) of the surface and subsurface soils that form restrictive or semi-restrictive boundaries to vertical water movement (Klaus & Jackson, 2018; Whipkey, 1965). The contrast in K values does not have to be abrupt, and in fact, gradual changes are more typical in soils (Whipkey & Kirkby, 1978). Generally, soil hydraulic conductivity decreases with depth, often exponentially (Ameli et al., 2016; Beven, 1982; Elsenbeer et al., 1992; Pirastru et al., 2017), and this decrease with depth can generate interflow (Zaslavsky & Sinai, 1981a). Interflow initiation and movement can be rapid, contributing to storm responses, and can persist for hours, days, or even weeks and hence also contribute to baseflow.

Definitions and conceptualizations of interflow have varied over time and amongst investigators, sometimes reflecting differences in the hillslope environments being investigated (e.g., Figure 1). Interflow exhibits threshold behaviour such that a combination of rainfall or

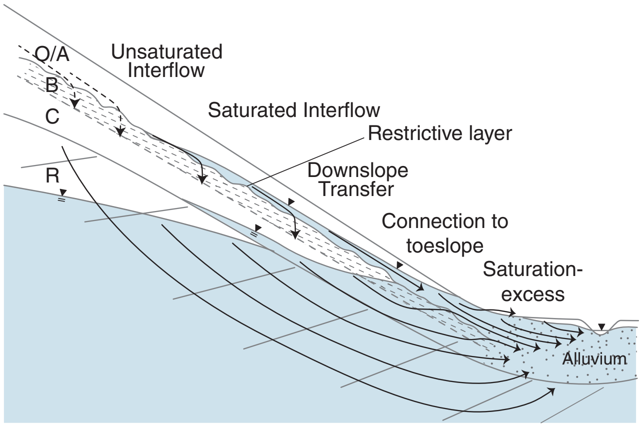

FIGURE 1 Conceptualization of percolation-limited (see upper slope) and transmissivity feedback (see toeslope) interflow and the merging of these interflow processes with groundwater flow and saturation excess streamflow generation. Percolation limitation interflow (see Section 2) can occur as both unsaturated and saturated flow whenever downward-moving soil water reaches a low-conductivity restrictive layer (e.g., a lower hydraulic conductivity B horizon) faster than it can percolate into the restrictive layer. Transmissivity feedback interflow occurs when the water table rises into higher conductivity soil or rock near the surface. The flowlines illustrated through the shallow fractured and weathered bedrock (i.e., depicted by the hashed lines in the R layer) between the soil and unweathered bedrock below indicate another form of interflow (e.g., Anderson et al., 1997; Asano et al., 2023).

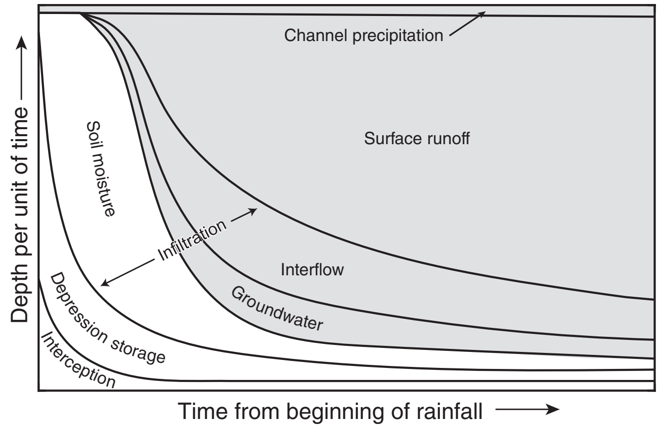

FIGURE 2 Early conceptualization of how interflow fits into the temporal variation of stormflow generation adapted from Linsley et al. (1949).

snowmelt and antecedent soil moisture are required for the wetting front to reach the restrictive layer and initiate downslope flow as the percolation rate exceeds the leakage rate through the restrictive layer (Jackson et al., 2014; Jackson et al., 2016; Kampf, 2011; Tromp-van Meerveld & McDonnell, 2006a, 2006b). Interflow can also occur as unsaturated flow (Figure 1; Gannon et al., 2017; Harr, 1977; Hewlett, 1961a, 1961b; Hewlett & Hibbert, 1963; Zaslavsky & Sinai, 1981b), but field studies usually focus on saturated interflow simply due to the difficulties of installing a dense enough network of sensors to determine soil water pressure head gradients, and because interflow rates at saturation are often orders of magnitude higher than soils below saturation.

Confusingly, three different terms have been applied to these hillslope flow processes: subsurface stormflow (e.g., Hursh, 1936; Lowdermilk, 1934; Weiler et al., 2006; Wisler & Brater, 1949), interflow (Barnes, 1942; Beven, 1989; Linsley et al., 1949) and throughflow (e.g., Kirkby & Chorley, 1967; Knapp, 1974) with no meaningful distinction amongst the terms (McMillan, 2022). As an example, two side-by-side chapters from the same book (Arnett, 1974; Knapp, 1974) use interflow and throughflow to refer to the same subject. Subsurface stormflow has sometimes been used as a catch-all term for all processes that together produce stormflow not explained by saturation overland flow (Dunne & Black, 1970) or Horton overland flow (Dunne et al., 1975; Hursh & Brater, 1941). This might include groundwater ridging as well as any other subsurface process that contributes to stormflow (Sklash & Farvolden, 1979; Weiler et al., 2006). Certain investigators and research groups have exclusively used one term over time, but other investigators have published using different terms synonymously in different papers and even in the same paper! It seems unlikely at this point that the community will settle on one term. The concept of interflow, although not always explicitly labelled as such, appears early in hydrology texts and literature. In this early documentation, interflow is described by various terms, including subsurface stormflow (Hursh, 1936), lateral intrasoil flow (Rodnikov, 1940 [translated]), shallow seepage of storm runoff (Lowdermilk, 1934) and the phenomenon of water that "reaches the headwater streams by exfiltrating from the soil, either by flowing above the impermeable bedrock or through larger and smaller veins". (Engler, 1919 [translated]). Bunting (1964) refers to hollows or convergent topographic regions as peroclines, which are "sites of moisture accumulation and lateral downslope movement of moisture". Interflow was presented as a major stormflow generation process in the 1949 textbook, Applied Hydrology (Linsley et al., 1949; Figure 2).

2 | MECHANISMS AND CONTROLS OF INTERFLOW

Interflow occurs as both saturated and unsaturated flow and can be produced by three main mechanisms: (1) percolation limitation when the downward flow of water in a soil exceeds the rate at which it can percolate into the next layer, akin to an infiltration-excess condition, (2) transmissivity feedback when the water table rises into higher conductivity soils near the surface and (3) downslope flow due to an anisotropic hydraulic conductivity field. Different mechanisms can be operating in different parts of the same hillslope. In the first two cases, the occurrence of interflow is predicated upon a decrease in hydraulic conductivity with depth resulting from changes in bulk density, texture, porosity and structure due to pedogenic processes or lithologic contact.

Soil pedogenesis and mineral weathering create near-surface hydraulic conductivity contrasts in many hillslopes. As soils weather, small particles (clays), soluble and colloidal iron, various salts and organic compounds are preferentially transported downward in the soil profile by infiltrating water, so older soils often develop an illuviated B horizon with accumulations of either clays, iron or aluminium oxides, or organic compounds (Chittleborough, 1992; McDaniel et al., 2001; Quénard et al., 2011). Hydraulic conductivities of these B horizons are typically much lower than the conductivities of the more biologically active topsoils (A, E and transitional horizons) in which roots and animals create high porosities and macropores. More developed B horizons such as argillic, placic, natric, petrocalcic, or plinthic horizons or fragipans and duripans (Soil Survey Staff, 1999) create more abrupt hydraulic conductivity contrasts leading to greater restriction of vertical water movement. Hydrological processes not only drive the translocation and accumulation of clays, which can develop argillic horizons and conditions for interflow via perched saturation, but perched saturation and interflow are also thought to degrade these argillic horizons (Jamagne, 1984). Soil development processes also occur laterally downslope providing evidence of subsurface transported material from interflow processes (Bourgault et al., 2017; Jackson, 1965; McDaniel et al., 1992; Park & Burt, 1999; Pilgrim et al., 1978). Interflow is also thought to differentiate soils in a catena or toposequence (Bailey et al., 2014; Zaslavsky & Rogowski, 1969).

When infiltrating water reaches restrictive layers faster than it can percolate through them, flow lines above the restrictive layer orient downslope, creating downslope and normal flow vectors (Zaslavsky & Sinai, 1981b). Simplistically, this partition is determined by the relative size of the downslope Buckingham–Darcy flux ([Buckingham, 1907]; K of the upper layer times $\sin\theta$, where θ is the slope angle) to the normal Buckingham–Darcy flux, which can be simplified to K of the lower layer times a unit hydraulic gradient. The same partition is defined by the relative Buckingham–Darcy fluxes when the flow is unsaturated. Under drier soil conditions, the unsaturated hydraulic conductivity of the lower layer is often higher than that of the topsoil, and this is why it is necessary for percolating water to wet a porous or fractured restrictive layer before interflow initiates (Hopp et al., 2011; Hübner et al., 2017).

If the contrast between the infiltration and percolation rates is high, a saturated wedge forms above (and within) the restrictive zone (Jackson et al., 2014; Weyman, 1973; Whipkey, 1965), and this wedge of saturation can be perched (episaturation) where the soil or rock below the restrictive layer is below saturation (Figure 1). When the infiltration and percolation rate contrast is lower, interflow occurs as unsaturated downslope flow if the capillary-pressure head dependent hydraulic conductivity, $K(\psi)$, of the upper layer is higher than the $K(\psi)$ of the lower layer as illustrated by simulations of Zaslavsky and Sinai (1981b). Whether this is significant in the field is still a matter of debate. Hewlett (1961a, 1961b), Hewlett and Hibbert (1963) and Lee et al. (2020) examined long-term drainage in 7 and 15 m long inclined soil models containing repacked native soils starting at near saturated conditions and found that the saturated wedge quickly contracted, but that unsaturated flow from upslope continued to feed the saturated zone at the base of the slope. Harr (1977) found that unsaturated interflow was dominant in all but the lower 12–15 m of a mountain slope in Oregon. Observations and Richards (1931) equation modelling of snow-dominated semi-arid slopes also indicate substantial unsaturated interflow feeding valleys (McNamara et al., 2005). By its nature, unsaturated interflow is hard to detect and quantify, and this may explain the scarcity of field studies investigating interflow in under-saturated conditions. Saturated interflow can be equally challenging to observe when the saturated thickness over the restrictive layer is thin or in dendritic networks (Graham, Woods, & McDonnell, 2010; McGuire & McDonnell, 2010).

Saturation of near-surface high-conductivity soil layers can also occur as a bottom–up process from a water table that rises upward into a more transmissive layer as the deeper storage fills. This rising water table reaches the dense macropore network of the near-surface environment in which the bulk hydraulic conductivity may be orders of magnitude greater (Beven & Germann, 2013), and the near-surface water moves rapidly downslope (Figure 1; Bishop et al., 2011; Detty & McGuire, 2010; Kendall et al., 1999; Laudon et al., 2011; Lundin, 1982; Pirastru et al., 2017). This process is sometimes referred to as the transmissivity feedback mechanism (Bishop et al., 2011; Kendall et al., 1999) and can occur anywhere that topography, lithology, weather and water table dynamics bring the water table near the surface, but it is most common at the toeslope.

As reviewed by Beven and Germann (1982, 2013), macropore, pipe-flow and preferential flow have large effects on flow and solute transport in the near-surface environment where roots and animals create higher densities of macropores, many of which spread laterally (e.g., Ghestem et al., 2011). These macropores, typically defined as a pore exceeding 0.6 mm diameter (Beven & Germann, 1982), when saturated allow some water to bypass matrix flow and hasten the movement of water and solutes down gradient. Macropore flow can also occur through unsaturated soil (Nimmo, 2011). In some hillslopes, macropore flow dominates interflow processes (e.g., Beasley, 1976; Mosley, 1979; Roberge & Plamondon, 1987; Uchida et al., 2005) and may serve to reduce soil pore pressures and thus lessen the chance of landslides on steep slopes (Uchida et al., 2001). Preferential flow has been found through tracer studies even where macropore flow is not observed (e.g., Hornberger et al., 1991; Jackson et al., 2016; Jarvis et al., 2016). As macropores are more common in topsoils and shallow soils, macropores increase the bulk conductivity of the surface soils and thus promote interflow over the lower conductivity impeding layers below (e.g., Beasley, 1976). De Vries and Chow (1978) even noted that wetting of the soil matrix occurs not only through vertical wetting front during rainfall, but also internally by the wetting of the periphery of macropores. Macropore and pipe flow both exhibit threshold behaviours with such flow becoming important in wet conditions (Uchida et al., 2001, 2005). Capturing the effects of macropores, soil pipes and preferential flow paths on the bulk conductivity requires large-scale conductivity measurements, such as irrigation experiments, or a high number of small-scale conductivity measurements. Soil macropores and preferential flow are considered to be ubiquitous (Clothier et al., 2007; Jarvis, 2007).

Theoretically, anisotropic soils (e.g., Assouline & Or, 2006; Bathke & Cassel, 1991; Beckwith et al., 2003; Soracco et al., 2010) can produce interflow, but the authors know of no field studies clearly connecting anisotropy in near-surface soils with interflow generation. McCord and Stephens (1987) found downslope movement of salt tracers in an unlayered sand dune and hypothesised soil anisotropy as the cause. Subsequently, McCord et al. (1991) conducted modelling that indicated state-dependent anisotropy could explain the downslope salt movement in an unlayered sand dune. State-dependent anisotropy is created when the moisture front creates a higher hydraulic conductivity layer above the dry and consequently lower hydraulic conductivity layer below the wetting front. However, Jackson (1992) demonstrated with a Richards equation model that on any sloping soil surface, the switch from a flux boundary condition during rainfall to a no-flux condition after rainfall would turn the near-surface flow vectors downslope and parallel to the slope (see also Sinai & Dirksen, 2006; Weyman, 1973). When the soil dries and the surface becomes an evaporative boundary, this creates conditions where part of the flow field also orients downslope, and either of these conditions could explain the downslope movement of a tracer in an unlayered hillslope. This boundary condition driven downslope flow is interesting from a solute transport viewpoint but is unlikely to move significant quantities of water downslope since such boundary-driven unsaturated fluxes are quite small during under-saturated soil conditions.

The interflow system at any point on the hillslope has a maximum capacity based on the soil thickness above the restrictive layer, the surficial lateral conductivity and the slope. When this capacity is exceeded, some interflow leaves the system to travel as saturation-excess flow (Beven & Kirkby, 1979; Dunne & Black, 1970). In such cases, the saturated interflow wedge fills the entire soil profile to the ground surface, creating variable source areas (near stream) and return flow or exfiltration high on the hillslopes and disconnected from the surficial water table (Jackson & Cundy, 1992; Sidle, 1984; Sidle & Tsuboyama, 1992; Torres et al., 1998).

3 | PORE-WATER VELOCITIES AND CELERITIES

Interflow is important for the advective movement of solutes and the stormflow response of streams, but the timescale of these two processes differs greatly in systems that are saturated or near-saturated due to the difference between the pore-water velocity and the celerity of pressure transmission through the interflow system (McDonnell & Beven, 2014). Celerity refers to the speed with which a wave propagates through the flow domain whereas the velocity of pore water describes the advective movement of water and controls the transport of tracer or solute through porous media. Rapid celerities of interflow partly explain the dominance of old-water discharge from the bottoms of slopes even after substantial quantities of new rainfall have fallen (Beven, 1982; Germann et al., 1986; Williams et al., 2002). This downslope propagation of pressure pushes or displaces downslope water rapidly for soils that are saturated or near-saturated (Rasmussen et al., 2000). To understand transport of water and solutes through catchments, and to build better-informed and more realistic hydrologic models, it is imperative that field studies simultaneously collect both hydrometric data (i.e., flows, water levels, soil moisture content) and tracer transport data to provide information on pore-water velocities.

4 | INITIATION AND FREQUENCY OF INTERFLOW

The initiation of interflow via saturated flow is controlled by processes that lead to saturated conditions in hillslope soils, provide continuity of those saturated conditions along a hillslope and allow for subsurface flow along the slope. The apparent effect of these processes is that interflow initiation exhibits threshold behaviour due to a series of storages that must be satisfied such as interception storage (Figure 3a,b), wetting front propagation (Figure 3c), the filling of subsurface depression storage (Figure 3d) and downslope wetting (Figure 3e). The vertical movement of water is often rapid and aided by preferential pathways (Beven & Germann, 2013). Wetting front migration and advective transport of water vertically through soils to depths where restriction might occur can be as rapid as a few millimetre per hour for advective transport and up to 10–15 times higher for wetting front movement or pressure wave propagation, which translates through the subsurface as the celerity response rate (Torres et al., 1998; van Verseveld et al., 2017). Once soils are saturated and hydraulically well-connected, a wave can propagate downslope yielding a rapid interflow response at the bottom of the hillslope (Figure 3e; Germann et al., 1986).

A rising water table, especially in lower slope positions, can drive a rapid increase in interflow as the saturated zone intersects soils with higher macroporosity (Figure 3f, i.e., the transmissivity feedback mechanism discussed earlier). Interflow may also contribute to and interact with groundwater ridging near the stream (Figure 3g). The capillary-fringe groundwater ridging hypothesis may also induce interflow and increase streamflow contributions from the shallow subsurface (Gillham, 1984; Sklash & Farvolden, 1979). This mechanism may occur where a surficial aquifer is near the ground surface and the capillary fringe (i.e., a zone of tension saturation) occupies most of the available soil storage. In such a case, only a small amount of water is needed to convert the storage from tension saturation to positive pressure causing a rapid change in the water table during storms creating a “ridge” of higher hydraulic head that facilitates increased discharge of riparian groundwater to the stream. The groundwater ridge also causes a temporary reversal of hydraulic gradients away from the stream during the early development of the ridge until all storage is filled and the hydraulic gradient is directed towards the stream (Cloke et al., 2006; Vidon, 2012). The importance of groundwater ridging as well as the topographic, rainfall and soil conditions necessary for its formation (Katsura et al., 2014) are both controversial (Buttle & Sami, 1992; Cloke et al., 2006; McDonnell & Buttle, 1998;

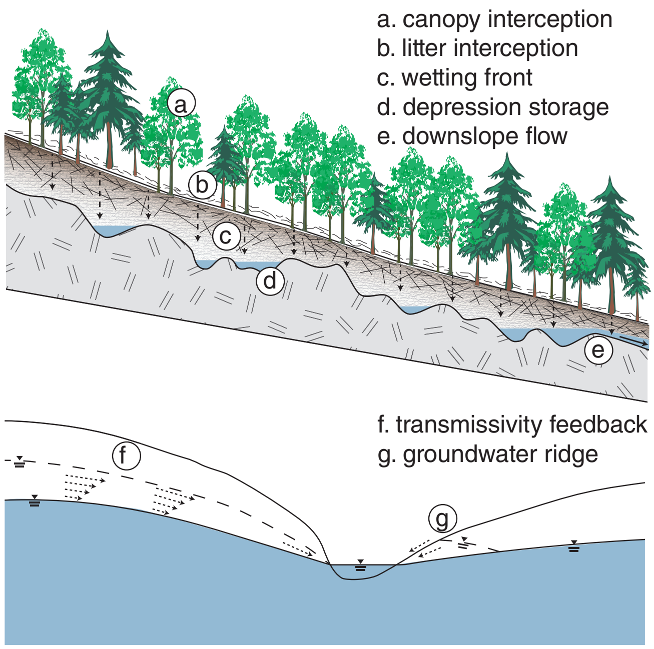

FIGURE 3 Top: top–down thresholds that must be met for rainfall to initiate interflow include (a) canopy interception, (b) litter interception, (c) water necessary for the wetting front to reach the restrictive layer, (d) water necessary to fill concavities in the restrictive layer, which can be a leaky restrictive layer of soil, saprolite, or bedrock and (e) water necessary for the saturated wedge to reach the near-surface macropore system and flow downslope. Bottom: bottom–up thresholds include (f) water table rise into more transmissive soil inducing macropore flow, and (g) groundwater ridging where the hydraulic gradient can rapidly increase forming a "ridge" through the conversion of a capillary fringe to a saturated zone.

Szilagyi, 2006; Waswa & Lorentz, 2015). Furthermore, observational evidence for groundwater ridging is again limited by the density of pressure head measurements necessary to characterise its occurrence. In any case, interflow delivers water to the toeslopes and valley margins where vadose zone storage is minimal, and recharge is rapid.

The time for saturation to develop at the restrictive layer is dependent on the interaction between the rainfall forcing, antecedent soil moisture and soil and topographic properties such as the depth to the restriction layer, the hydraulic conductivity difference between the restricting layer and overlying soil layers, and surface and subsurface topographic convergence. Once saturation forms, hydraulic connection is necessary to produce interflow, which is contingent on filling depressional storage of the restriction layer surface (Figure 3d; Ali et al., 2011; Buttle et al., 2004; Tromp-van Meerveld & McDonnell, 2006a, 2006b). Lateral hydraulic conductivity, leakage into the restrictive layer, and slope morphology such as length, gradient and curvature affect pressure wave propagation to the bottom of a slope. The effect of these controls on the interflow response time often leads to an apparent threshold-like response of interflow initiation to rainfall amount or intensity and antecedent moisture (Buttle & Turcotte, 1999; Tromp-van Meerveld & McDonnell, 2006a, 2006b). Studies have reported a large range of thresholds (18–60 mm) to storm precipitation (e.g., Du et al., 2016; Mosley, 1979; Tani, 1997; Uchida et al., 2005; Weiler et al., 2006) and indeed there are similar thresholds for runoff response at the catchment scale that are driven by processes other than interflow initiation (e.g., Ali et al., 2013; Ross et al., 2021). However, the influence of antecedent moisture and precipitation intensity on interflow response and filling storage is not easily generalised. It is dependent on how interflow is formed at a particular site, which is affected by soil structure and hydrologic properties such as available soil water storage and water retention properties above a restrictive layer and contrasts in hydraulic conductivity between soil layers and bedrock (Beven, 1982; Dusek & Vogel, 2016; Kienzler & Naef, 2008).

5 | DURATION OF INTERFLOW

Interflow from hillslopes can provide substantial amounts of stormflow to the stream once activated from precipitation or snowmelt events (e.g., McGlynn & McDonnell, 2003). The duration of interflow is a function of the duration, amount and intensity of the precipitation or snowmelt event and hillslope properties such as hydraulic conductivity, soil moisture characteristic parameters, depth and slope angle (Beven, 1982; Hopp & McDonnell, 2009). Hillslope and soil trenches provide valuable insights on interflow duration, even if interflow at a trench ceases earlier than uninterrupted interflow due to pressure conditions at the trench surface (Atkinson, 1978). A few studies on hillslope trenches reported persistent interflow over weeks to months in snowmelt driven systems (Kim et al., 2004; Ohara et al., 2011). However, interflow persistence after storm events varies widely as a function of soil material and slopes ranging from a few hours to days (e.g., Freer et al., 2002; Woods & Rowe, 1996). Rinderer et al. (2021) showed at various trenched forested hillslopes in southern Germany that the perching duration was consistent between seasons and ceased within one day after peak flow. Dahlke et al. (2012) and Jost et al. (2012) found rather short recession times of a few hours. In the classic hillslope drainage study at the Coweeta Hydrologic Laboratory, Hewlett and Hibbert (1963) found that interflow continued to contribute to the downslope drainage for 145 days, however on a concrete foundation. Lee et al. (2020) repeated the Hewlett and Hibbert experiment 50 years later and found that the drainage only lasted for about 14 days and attributed the difference to leakage from the concrete bottom of the experimental trough and suggested that those results might reflect similar drainage times from real hillslopes where there is vertical leakage of interflow into the restrictive layer. In a 407 mm, 52 h-irrigation experiment, Jackson et al. (2016) observed that a perched water table persisted above a trench for about 60 h after irrigation stopped, and trench outflow continued for an additional day.

6 | TRAVEL DISTANCES AND CONNECTIVITY

Perched interflow behaves similarly to Horton overland flow in several ways (McDonnell, 2013). First, it occurs when water arrives at a surface faster than it can infiltrate or percolate into that surface. Second, it is affected by depression storage (i.e., the "fill and spill" behaviour noted on several hillslopes, e.g., Buttle et al., 2004; Du et al., 2016; Jackson et al., 2016; Nyquist et al., 2018; Tromp-van Meerveld & McDonnell, 2006a, 2006b; and in a commentary by McDonnell et al., 2021). Third and most importantly, downslope lateral flow can occur whilst some soil water continues to percolate through the impeding layer below (Ameli et al., 2015; Lee et al., 2020; McDonnell, 2013). However, field investigations and modelling efforts of interflow have usually assumed that percolation through the restrictive layer was negligible because of the high contrast in hydraulic conductivities across the restrictive layer, without measuring the conductivity of the lower layer or the percolation rate through it. This assumption is often extended to assuming that all saturated interflow, perched or otherwise, would connect directly to the valley or stream (e.g., Detty & McGuire, 2010; Jencso et al., 2010).

Over the last two decades, evidence has emerged that percolation from perched interflow episodes can be substantial and reduce or eliminate connectivity between interflow-producing slopes and stream valleys. Even where interflow occurs when water tables rise into high conductivity surficial layers, the water table may drop below this layer not long after rainfall. Several irrigation experiments on soils overlying bedrock (Graham, Verseveld, et al., 2010; Tromp-van Meerveld et al., 2007) or high clay content Bt horizons (Jackson et al., 2016) have demonstrated substantial percolation losses simultaneous with interflow occurrence, challenging the negligibility assumption. Furthermore, Du et al. (2016) found that Darcian travel times from hillslopes to streams were much longer than perched interflow events and concluded interflow from most of the hillslopes could not reach valleys or streams. To assess this issue of interflow connectivity (cf. Wilson et al., 2017), Jackson et al. (2014) identified a simplifying assumption for analysing perched interflow, that the pressure head was zero at the bottom of the restrictive layer, and this assumption allowed the calculation of downslope travel distances for saturated interflow based on vector addition, like earlier direct analytical equations for predicting interflow direction (Ahuja & Ross, 1983; Zaslavsky & Rogowski, 1969). Travel distance represents the length that a water molecule traverses in the saturated zone; thus, it is related to the Darcy flux not its celerity. This simple geometric model reveals that the downslope travel distance before percolation into the restrictive layer of water moving downslope in a perched saturated wedge is (Jackson et al., 2014):

$$L_D = (K_u/K_L)(\sin(\theta)/[(N+C_n)/C_n])N, \tag{1}$$

where (K_u/K_L) is the hydraulic conductivity ratio between the upper (u) and lower (L, restrictive) layers, N is the thickness of the perched saturated wedge above the restrictive layer, $\sin\theta$ is an approximation for the hydraulic gradient in the downslope direction, and C_n is the thickness of the restrictive layer. The gradient ratio, that is, the second parenthetical term in Equation 1, depends on the thickness of the restrictive layer, which is often not known, but as the thickness increases, downslope travel distance increases, reaching a limit as follows:

$$L_D = (K_u/K_L)\sin(\theta)N. \tag{2}$$

As all field studies of interflow have indicated, this equation shows that interflow is more important on steep slopes with a large ratio between the conductivities of the interflow layer and the restrictive layer. Ahuja and Ross (1983) similarly predicted downslope travel distances based on the conductivity ratios and slopes. They found with any given rate of leakage into the restrictive layer, even small amounts of leakage, interflow travel distance was significantly reduced compared to interflow over an impermeable base. It is important to note that the travel distance estimate in Equation 1 applies only where the interflow layer is perched and the lower layer is not under positive pressure, which is likely common in upland settings and not near the stream. Downslope travel distances vary across three orders of magnitude for common hillslope arrangements (Figure 4) and exist in a continuum of importance as a streamflow generation mechanism.

Klaus and Jackson (2018) subsequently conducted a meta-analysis of 17 hillslopes with sufficient published data to calculate travel

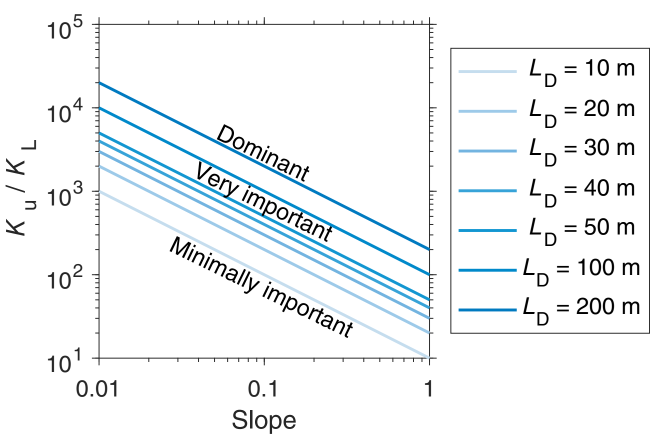

FIGURE 4 Perched interflow travel distances, L_D, as a function of slope, $\sin\theta$ and the ratio of the hydraulic conductivities (K_u/K_L) of the upper layer and the restrictive layer. Greater travel distance suggests a greater importance of interflow as a streamflow generation mechanism.

distances, and travel distances varied from less than a metre to several hundred metres across these sites. Evaluating the absolute distances against slope lengths, travel distances were less than 50% of slope distances at 14 sites and less than 30% at 11 sites. In 14 of 17 cases, most water perched above a shallow restrictive layer percolates through the restrictive layer before reaching the valley. In such cases, interflow would contribute directly to the valley only from the lower portions of the hillslopes. Percolation through the restrictive layer cannot be ignored when evaluating hillslope to valley connectivity.

Topsoils feature dense networks of macropores created by roots and various soil-dwelling animals including ants, earthworms and even vertebrates such as salamanders and moles (Aubertin, 1971). These macropores convey water even when the soil is unsaturated, but conveyance is enhanced when the soil saturates (Beven & Germann, 1982, 2013), contributing to the transmissivity feedback process discussed above (Figure 3f). Many studies have noted and measured tracer velocities in interflow that cannot be explained by the bulk saturated conductivity of the soil (e.g., Buttle & McDonald, 2002; Hammermeister et al., 1982; Leaney et al., 1993; McGlynn et al., 2002). Some investigators have noted apparent rapid preferential flow in the absence of visible macropores, so the term preferential flow has been applied to rapid flow that cannot be explained by matrix flow (Gerke, 2006). Tracing studies have shown that solutes moving through the macropores can move at pore-water velocities orders of magnitude greater than those in the matrix (e.g., Jackson et al., 2016). Correspondingly, we can think of two downslope travel distances—the travel distance for matrix flow and the travel distance for macropore flow, which will be much longer. This also would be the case when macropore networks are discontinuous. Pressure wave propagation downslope in the soil matrix may connect macropores such that the travel time is a combination of water transport through macropores and displacement of stored water (Beven & Germann, 2013; McDonnell, 1990).

The downslope travel distance formula applies to perched interflow as the assumption of zero pressure head at the base of the restrictive layer does not apply when the surficial water table merges with the interflow layer. However, if the hydraulic gradient in the impeding layer is downward or nearly so, the equation still provides an approximation of downslope travel distance. In the case where groundwater is moving into the interflow layer from below, travel distances become unbounded and connectivity to the toeslope or valley can be assumed.

From ridge to valley, hillslopes vary in gradient and shape, from concave upwards to downwards, and thus interflow travel distances also vary from ridge to valley. In older topography, where ridgetops are relatively flat, perching above restrictive layers may occur, but hydrologic fluxes are still largely vertical. In mid-slopes, interflow may transfer water and solutes downslope but will still percolate to the water table before reaching the valley (Jackson et al., 2014). Only in steep mountainous topography with strong conductivity ratios is interflow likely to connect water infiltrating near the ridge to the valley (Klaus & Jackson, 2018).

The flow and solute connectivity of interflow to streams depends strongly on valley morphology. If the stream is located at the foot of the hillslope, then interflow can pass directly from the hillslope to the stream, contributing directly to stormflow or the recession limb of the hydrograph and preserving the chemical signature of the interflow input. Such a situation is uncommon, however. More typically interflow delivers water to the riparian aquifer, and consequently the effect on both stormflow and stream chemistry is buffered by mixing with hillslope and riparian groundwater (e.g., Benettin et al., 2015; Hill, 2000; Jencso et al., 2010; McGlynn et al., 1999; McGlynn et al., 2003; Mulholland, 1992; Ploum et al., 2020). If the surficial aquifer and the interflow wedge are merged at the toeslope, their chemical signatures and hydrograph effects will be mixed, and some of the flow may become return flow and saturation-excess runoff before reaching the stream. The toeslope and the riparian aquifer are the locations where many streamflow generation processes merge together, complicating the interpretation of hydrograph components and stream chemistry.

7 | IMPLICATIONS AND IMPORTANCE

A variety of observational approaches (e.g., Blume & Van Meerveld, 2015) can be applied to the question—does the occurrence of interflow matter at the catchment scale? Tracer-based hydrograph separation has been applied in many catchments around the world to determine stormflow sources for hillslopes and catchments. While there is a wide range of approaches on how to sample sources for determining their contributions (Buttle, 1994; Klaus & McDonnell, 2013), there is a body of tracer-based results suggesting that one can detect substantive contributions from interflow at the catchment outlet during events. Terminology varies between tracer-based studies, for example, referring to "soil water" (Ogunkoya & Jenkins, 1993), "old hillslope water" (McGlynn & McDonnell, 2003), "shallow subsurface flow" (Hogan & Blum, 2003), or "subsurface stormflow" (DeWalle & Pionke, 1994). For a spring located in a toeslope position, Iwagami et al. (2010) showed that "soil water" is a major source of flow. At the Maimai catchment, where interflow is a dominant process (Gabrielli & McDonnell, 2020; Klaus & Jackson, 2018; McDonnell et al., 1991), "old hillslope water" can contribute up to 50% of streamflow during storms at the scale of a small catchment. At Hubbard Brook, a tracer-based study revealed contributions from "shallow subsurface flow" (equated to soil water) on the order of 40% of runoff at the catchment outlet during events. Studies have found "soil water" contributions of 50% and even greater than 75% in steep catchments in the eastern USA (Bazemore et al., 1994; McHale et al., 2002; Peters et al., 1995). In a nested catchment study, James and Roulet (2009) used a tracer approach to estimate contributions from a "perched water table" to storm runoff. The contributions varied widely between catchments and storms, mostly ranging from 10% to 30% but up to 57%. Some of this variation may be due to differences in transport compared to the pressure wave displacement of old water (Beven & Germann, 2013; Kienzler & Naef, 2008). In general, sites with notable slope gradients and shallow soils tend to favour greater contributions of soil water to catchment event runoff (Carey & Quinton, 2005; DeWalle & Pionke, 1994; Hinton et al., 1994; Rice & Hornberger, 1998).

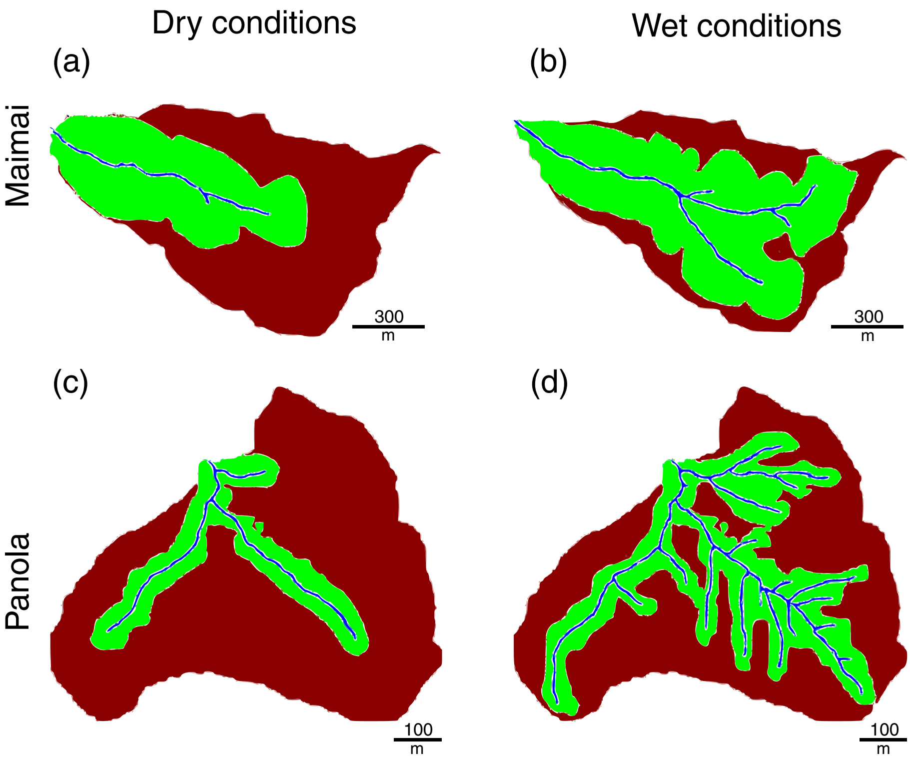

FIGURE 5 Conceptualization of contributing area (green) for wet and dry conditions at the Maimai M8 (top, (a) & (b)) and Panola (bottom, (c) & (d)) catchments that connects hillslope interflow to the stream depending on the extension of the stream network. The contributing area is estimated using downslope travel distances determined by Equation 2, the digital terrain model for each site, and assuming homogeneous subsurface characteristics for each site. As the stream network expands, so does the contributing area, but the expansion is greater at Maimai because it has steeper slopes and a greater hydraulic conductivity contrast between the restrictive layer and the overlying soil.

The hydrological community struggles to generalise observational findings across landscapes. Without understanding the landscape controls on interflow generation and connectivity to valleys, we cannot interpret streamflow behaviour and solute transport. In landscapes with slopes over 7% and shallow layers that impede percolation, perched interflow can be a substantial contributor to stormflow and can support baseflow as well. In flat landscapes where the water table frequently rises into high conductivity surface soils, the resulting transmissivity feedback can also move large amounts of water laterally in the near-surface environment and contribute to stormflows. Interflow can also laterally redistribute water from higher to lower on hillslopes, even when interflow travel distances are short. This lateral redistribution of soil water has implications for vegetation (e.g., Hwang et al., 2012) and can improve recharge estimates in groundwater models (Staudinger et al., 2019).

Interflow, in all its forms, acts as a subsurface extension to variable source areas (Ameli et al., 2015; McDonnell, 2013). As a topographically-driven process, interflow delivers water precisely to the areas where variable source areas tend to form (toeslopes, hollows, channel and wetland margins and floodplains) (e.g., Anderson & Burt, 1978; Dunne et al., 1975; Glaser et al., 2020). We tend to think of variable source areas as forming where the surficial water table rises to the surface, but interflow often contributes to this rise and partly controls the dynamics of variable source areas. The observation of interflow delivering water to the base of slopes was fundamental to early ideas on variable source area behaviour (Hewlett & Hibbert, 1967). The catchment scale importance of interflow is particularly influenced by the extension and contraction of the stream network as it affects the effective pathlength connecting interflow source areas that contribute to the stream network (Pardo et al., 2022; van Meerveld et al., 2019). In Figure 5, we conceptualise contributing areas via interflow through a shallow saturated layer in the Maimai M8 (McGlynn et al., 2002) and Panola (Aulenbach et al., 2021) catchments as examples assuming homogeneous subsurface characteristics. The extent of this contributing area is controlled by contrasting downslope travel distances between both sites (Klaus & Jackson, 2018) due to the subsurface and topographic characteristics as well as differences in stream network density that would vary due to catchment wetness conditions. The steeper slopes and greater hydraulic conductivity contrast at Maimai compared to the Panola catchment suggest greater travel distances at Maimai. As the stream network expands with increasing wetness, the contributing area at Maimai nearly occupies the entire catchment.

8 | NUMERICAL MODELLING

Numerical models have been used to evaluate the role of interflow on streamflow generation, but it remains questionable how accurately hydrologic models predict interflow occurrence, flux, spatial variability and connectivity (Beven, 2001). It is a tenet of hydrologic modelling that simulations will be more accurate and more robust if they realistically represent runoff processes, including interflow (see Grayson et al., 1992; Kirchner, 2006). It is also argued that Earth system models will strongly benefit by accounting for interflow (Fan et al., 2019). In practise, interflow representation in hydrologic modelling takes many forms including non-existent (e.g., large-scale land surface models like VIC; Liang et al., 1994), conceptual near-surface storage buckets (e.g., PRMS, HSPF and HBV; Singh, 2012), 1-D simplified Darcian flow using the kinematic wave or Boussinesq approximations without leakage through the restrictive layer (e.g., Beven, 1981; Sloan & Moore, 1984; Troch et al., 2003), TOPMODEL theory (e.g., Shaman et al., 2002; Walter et al., 2002), spatially-distributed Darcian flow (e.g., DHSVM; Wigmosta et al., 2002) and multi-dimensional physical modelling based on Richards equation (e.g., Ameli et al., 2015; Glaser et al., 2016; Sloan & Moore, 1984; Wilusz et al., 2020). In addition, models that include more detailed process representation might lead to more uncertain predictions or even the rejection of model structures given challenges associated with model parameterization, uniqueness of place and a common lack of available observations used to evaluate models (Beven, 2000).

Richards equation models have been used to examine discrepancies between field and laboratory observations of flow fields and model predictions, and these models have been used to conduct thought experiments or test hypotheses on interflow processes and controls (e.g., Beven, 1977; Ebel et al., 2008; James et al., 2010; Zaslavsky & Sinai, 1981b). Other representations of interflow are more conceptual rather than physical. Bucket models are an example where interflow is controlled by conceptual recession rates that must be calibrated (Crawford & Linsley, 1966; Markstrom et al., 2015; Seibert & Bergström, 2021). Recently, Bitew et al. (2020) developed and tested dynamic upslope boundary condition modelling based on the Darcy flux estimates of downslope travel distance described above. However, available spatial soil property databases rarely include sufficient data on soil hydraulic conductivities, moisture characteristic curves, or topsoil depth to accurately model interflow generation, interflow-producing parts of the landscape, or interflow connectivity to the valley without calibration.

9 | UNRESOLVED ISSUES—WHAT IS NEXT?

Hydrologists seek to predict dominant streamflow generation processes based on spatial information about topography, soils, geology and vegetation (e.g., Jehn et al., 2020; Scherrer & Naef, 2003; Schmocker-Fackel et al., 2007; Wolock et al., 2004) or even to infer them from hydrograph characteristics (e.g., Barnes, 1942; McMillan et al., 2022; Olden et al., 2012; Sawicz et al., 2011). With respect to interflow specifically, relevant questions are: (1) how do we efficiently characterise the soil, geologic and slope controls on interflow? (2) What rainfall/storage thresholds must be met to initiate interflow? (3) What is the distribution of interflow travel distances in a catchment? and (4) how does interflow interact with the water stored along its flowpath and other streamflow mechanisms merging at the toeslopes, particularly in resolving the differences between wave displacement of stored water and water transport from upslope? Finally, and most critically, (5) How can we answer these questions beyond individual site-specific data-intensive studies, but rather with readily available soil, geology, topographic and hydrograph data? Currently, it seems that some level of field investigation is necessary to satisfactorily answer these questions, which limits progress in catchment hydrology (Burt & McDonnell, 2015). Recent work using hydrologic signatures from continuous streamflow time series has shown that catchments can be classified into different dominant streamflow generation processes including interflow (Wu et al., 2021), but we still lack the ability to address the questions above given easily observable catchment characteristics.

Measuring interflow states and processes across catchments remains difficult. Whether interflow is perched locally somewhere in a hillslope can only be known if wells/piezometers or moisture/potential sensors are installed in the right places where interflow is occurring and deep enough to confirm the saturation state within and below the restrictive layer. Furthermore, fluxes and moisture dynamics in the vadose zone are generally not well-known in many places—even in intensively studied catchments. It is rare that observations are made at pertinent depths to understand perching dynamics. Furthermore, subsurface sensors installed in most research catchments and hillslopes are too few to reliably extrapolate water table positions or head dynamics across the catchment. Drilling wells in hard rock on many landscapes (e.g., steep, forested) is difficult, as is installing soil moisture sensors or tensiometers deeper than arm's length. Progress is being made on installing piezometers, moisture sensors and tensiometers in rock (Hahm et al., 2019; Rempe & Dietrich, 2018; Salve et al., 2012), but the field of hillslope hydrology still suffers due to the expense and impracticality of installing arrays of sensors and deep sensors in many hillslope environments. Geophysical techniques are bridging some of these gaps (e.g., Binley et al., 2015; Fan et al., 2019), but interpretation, expertise, spatio-temporal extent of the information and ground-truthing present challenges.

Accounting for the variability in flow processes, and the controls of those processes, across hillslopes and within catchments is a challenge. To predict interflow travel distances, numerous replicates of K measurements (e.g., Benton et al., 2022; Elsenbeer et al., 1992; Gabrielli et al., 2018) or large-scale irrigation experiments (e.g., Brooks et al., 2004; Graham, Verseveld, et al., 2010; Jackson et al., 2016; Markart et al., 2015; Tromp-van Meerveld et al., 2007; van Verseveld et al., 2017) are necessary to account for the high variability in conductivities, particularly in the near-surface environment. Working at Hubbard Brook Experimental Forest, Detty and McGuire (2010) collected limited conductivity data, which indicated the ratio of solum to C horizon hydraulic conductivities was only about two. Additional measurements from the same catchment indicated that the median conductivity ratio was about 130 (Benton et al., 2022), which increased the estimated downslope travel distance by over 50 times and changed the perception of interflow importance from minimal to dominant based only on Darcian estimates (i.e., without considering preferential flow) (Pardo et al., 2022). Many interflow studies have relied on data from a single interception trench of modest length (e.g., McGuire & McDonnell, 2010; Pilgrim et al., 1978; Tromp-van Meerveld & McDonnell, 2006a, 2006b; Weyman, 1973; Whipkey, 1965), but all studies that have used multiple trenches or long segmented trenches have found high variability in interflow production on the same hillslope (Du et al., 2016; Elsenbeer et al., 1992;

Woods & Rowe, 1996). This is true even in an artificial hillslope with soils designed to have uniform properties (Bauser et al., 2022).

Depressions in the restrictive layer are common (e.g., Du et al., 2016; Graham, Verseveld, et al., 2010; Kim et al., 2016; Phillips, 2018; Shouse & Phillips, 2016), whether it is composed of an argillic soil layer, glacial till, or shallow bedrock (e.g., Camporese et al., 2019; Du et al., 2016; Tromp-van Meerveld & McDonnell, 2006a, 2006b). The formation, frequency and depth of these depressions, however, are not well-explored, again because of technological issues. Depression topography and near-surface lithology can be explored manually using knocking poles, tile probes, or augers (e.g., Freer et al., 1997; Kim et al., 2016) but at high labour costs. Seismic refraction (Donaldson et al., 2023; Olyphant et al., 2016), electrical resistivity tomography (Gourdol et al., 2021; Kotikian et al., 2019; Tancredi et al., 2022; Thayer et al., 2018) and ground-penetrating radar (Nyquist et al., 2018) have all demonstrated some success in mapping soil and lithologic layers of differing hydraulic properties. All these techniques entail sampling challenges with respect to spatial resolution and characterising gradual transitions, rather than abrupt boundaries, in subsurface hydraulic properties. Advancing our understanding of the variability and controls of hillslope flow pathways will require increasing integration of geophysical technologies in hillslope hydrologic investigations.

Our understanding of interflow and the spatial variability of streamflow generation process comes from well-instrumented and intensively studied hillslopes and small catchments, but most modelling is done at larger scales using digital elevation models, land cover interpretation from satellite imagery, large-scale geologic maps and low-resolution, low-information soil maps. There is a large discrepancy from our complex conceptual models of streamflow generation processes and our lumped and averaged streamflow generation algorithms in our models.

10 | CONCLUSION

Experimental hillslope hydrology has advanced our understanding of the role of interflow in streamflow generation since the seminal work by Hewlett and Hibbert (1963). Today, the fundamental drivers of interflow are well understood, and a range of hydrological models account for interflow processes. Hydrologists have an excellent foundation to address the current gaps in understanding, simulation and generalisation of interflow across landscapes and scales. However, the community needs to overcome the often site-specific perspective on interflow processes and reverse the decline in the number of field studies in hydrology (Burt & McDonnell, 2015). Initiatives such as the Critical Zone Observatories provided great opportunities in understanding and generalising hydrological processes (Wlostowski et al., 2021). There is a clear community effort needed in re-vitalizing experimental hillslope studies especially with a view on synthesising across sites, as attempted two decades ago (McGuire et al., 2006; Retter et al., 2006) with consistent and transferable experimental protocols. Opportunities also exist through joint data initiatives in hillslope hydrology building on FAIR data sharing (Cudennec et al., 2020; Wilkinson et al., 2016), relying on meta-analysis (Evaristo & McDonnell, 2017) and involving machine learning to synthesise and generalise (Razavi et al., 2022). Opportunities are abundant for hydrologists using experimental or modelling approaches to improve our understanding of the often-elusive interflow process.

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DATA AVAILABILITY STATEMENT

No original data were used in this review.

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