Hydropedology and Surface/Subsurface Runoff Processes

HENRY LIN1, ERIN BROOKS2, PAUL Mc DANIEL3 AND JAN BOLL2

1Department of Crop and Soil Sciences, Pennsylvania State University, University Park, PA, US
2Department of Biological and Agricultural Engineering, University of Idaho, Moscow, ID, US
3Department of Plant, Soil, and Entomological Sciences, University of Idaho, Moscow, ID, US

The role of soils has long been recognized as critical to rainfall–runoff processes in watersheds. Hydropedology is an emerging interdisciplinary field that integrates pedology, hydrology, geomorphology, and other related bio- and geosciences to study interactive pedologic and hydrologic processes and the landscape–soil–hydrology relationships across space and time. This article presents an overview of hydropedology’s contributions to the understanding and modeling of surface/subsurface runoff processes, especially the diagnosis of soil features that can help answer “why-type” questions in watershed hydrology and the ubiquitous nature of preferential flow and its networks. We highlight two bottlenecks for advancing watershed hydrology and hydropedology: a conceptual bottleneck of modeling subsurface preferential flow networks and a technological bottleneck of nondestructively mapping or imaging subsurface architecture. Quantification of “soil architecture” at various scales and the identification of “hydropedologic functional units” in different landscapes offer promising potentials to advance hydrologic modeling. We present the information of linking surface/subsurface runoff processes to pedologic understanding at three scales of microscopic (macropores and aggregates), mesoscopic (horizons and pedons), and macroscopic (hillslopes and catchments) levels. Various examples from the literature are synthesized to illustrate the key points. Further research needs are suggested in the end.

INTRODUCTION

The role of soils (soil properties and their distribution over the landscape) has long been recognized as critical to rainfall–runoff processes in a watershed. Hydropedology, as an intertwined branch of soil science and hydrology, embraces multiscale studies of interactive pedologic and hydrologic processes and their properties in the variably saturated and unsaturated surface and subsurface environments, and therefore possesses significant potential in enhancing the understanding and prediction of rainfall–runoff processes (Lin, 2003; Lin et al., 2006a).

The traditional view of surface runoff generation has been based on the infiltration excess, or Hortonian, concept of runoff production (Horton, 1940), where surface runoff is the result of rainfall intensity exceeding infiltration capacity at the soil surface. This surface runoff, sometimes also called unsaturated overland flow, occurs more commonly in arid and semiarid regions, where rainfall intensities are high and soil infiltration capacity is reduced because of surface sealing or pavement. The primary motivation for many of the early studies was to determine the magnitude and return period of extreme events, as in many areas of the world these extreme events have been caused by infiltration-excess runoff.

An alternative runoff generation paradigm, called variable source area (VSA) hydrology, has been suggested over four decades ago (Hewlett and Hibbert, 1967). The VSA approach to watershed response suggests that runoff contributing to a storm hydrograph is generated from localized (typically near-stream) areas of the landscape. The controlling mechanism is termed saturation-excess runoff generation because of precipitation falling on soils that have little or no available storage (typically high soil moisture content or groundwater level), thereby precluding infiltration. This “saturated overland flow” occurs more frequently
in riparian zones, valley bottoms, or local depressional areas in humid or semihumid regions. As a consequence, watershed hydrologists must consider not only soil properties and their spatial distributions, but also the interactive roles of topography, land use/land cover, geology, and other components of the hydrologic cycle besides precipitation (Gburek et al., 2006).

Another runoff, called subsurface return flow (or reflow), occurs after water infiltrates the soil on an upslope portion of a hillslope and then flows laterally through the soil and exfiltrates (resurfaces) in the toeslope area or closer to stream channel (Whipkey, 1965; Dunne et al., 1975). This overland flow is related to subsurface stormflow (also called subsurface runoff, interflow, throughflow, lateral flow, transient groundwater, or some other names) (see Chapter 112, Subsurface Stormflow, Volume 3). Subsurface stormflow is another runoff-producing mechanism operating in most upland terrains, especially in humid environment and steep terrain with conductive soils or soils with a water-restricting layer. While some studies have documented subsurface stormflow as unsaturated flow in the vadose zone, most studies have shown that subsurface stormflow is saturated (or near-saturated) – due to either the rise of an existing water table into more transmissive soil above (with ensuing lateral flow) or the transient saturation above an impeding layer, soil–bedrock interface, or some zone of reduced permeability (e.g. fragipan, argillic horizon, and other dense layers in the soil profile).

Field investigations conducted in various humid forested catchments throughout the world have also revealed the significance of subsurface preferential flow at multiple spatial and temporal scales. While overland flow is limited in these forested catchments, evidence of subsurface preferential runoff is ubiquitous (Sidle et al., 2001; McDonnell, 2003; Lin, 2006). Growing evidence suggests that the subsurface flow network is a key to understanding rainfall–runoff processes including threshold behavior (Sidle et al., 2001; Weiler et al., 2005; McDonnell et al., 2007; Lin and Zhou, 2008), and that flow pathways often determine the hot spots and moments of biogeochemical processes (Band et al., 2001; McClain et al., 2003; Boyer et al., 2006). This suggests that a new thinking beyond the VSA paradigm is needed (Sidle et al., 2000; McDonnell, 2003; McDonnell et al., 2007). The dynamic origin of network structures in soils and hydrologic systems and recurrent patterns of self-organization are the subjects of recent research and model development (Rinaldo et al., 2006; James and Roulet, 2007; Weiler and McDonnell, 2007; Lehmann et al., 2007).

There are significant questions remaining in hillslope hydrology and rainfall–runoff prediction. For example, McDonnell et al. (2007) raised a number of such questions in their new vision of watershed hydrology: Why does heterogeneity exist? Why is there preferential, networklike flow at all scales? Why are hydrological connections at the hillslope and watershed scale so thresholdlike when the soil, climate, vegetation, and water appear so tightly coupled? This article attempts to shed light on some of these questions from a hydropedologic perspective. In particular, two bottlenecks are identified and discussed, where hydropedology can make a unique contribution to advancing watershed hydrology, including the quantification of soil architecture at various scales and the identification of hydropedologic functional units (HFUs) in different landscapes.

The article provides an overview of the current understanding of hydropedology’s role in rainfall–runoff processes, especially the diagnosis of soils that can help answer “why-type” questions in watershed hydrology. We illustrate various aspects that show the diagnostic importance of soils in rainfall–runoff processes at three scales – macroscopic (macro pores and aggregates), mesoscopic (horizons and pedons), and macroscopic (hillslopes and catchments) levels. Some research needs for the future are suggested in the end.

HYDROPEDOLOGY AND ITS FUNDAMENTAL QUESTIONS

Hydropedology, as an emerging interdisciplinary field, integrates pedology, hydrology, geomorphology, and other related bio- and geosciences to study interactive pedologic and hydrologic processes and the landscape–soil–hydrology relationships across space and time (Lin, 2003; Lin et al., 2005a, 2006a). It aims to understand pedologic controls on hydrologic processes and properties (e.g. soil as a central link in the hydrologic cycle and as a foundation for rainfall–runoff processes) and hydrologic impacts on soil formation, variability, and functions (e.g. hydrology as a critical driving force of soil formation and functional dynamics). Essentially, hydropedology seeks to answer the following two fundamental questions:

1. How the structure and distribution of soils over the landscape exert a first-order control on landscape water across spatiotemporal scales in the surface and the subsurface?
2. How landscape water (and associated transport of chemicals and energy by flowing water) impacts soil evolution, variability, and functions?

Landscape water here encompasses the source, storage, availability, flux, pathway, residence time, and spatiotemporal distribution of water (and the transport of chemicals and energy by flowing water) in the variably saturated and unsaturated surface and subsurface (Lin et al., 2006a). While source, storage, availability, and flux of water have been studied extensively in the past, attention to flow pathways (especially flow networks), residence time (age of water), and spatiotemporal pattern of flow dynamics
Hydropedology and Surface/Subsurface Runoff Processes

Soils (root zone)
Deep vadose zone (pores partially filled with water)
Ground water zone (pores entirely filled with water)
Hydrology
Hydrogeology

The pedon paradigm
Pedologic processes (soil)

The Critical Zone

The landscape paradigm
Geomorphologic processes (landscape)

Figure 1 Hydropedology as an interdisciplinary science that promotes integrated studies of hydrologic, pedologic, and geomorphic processes across spatial and temporal scales. It connects the pedon and landscape paradigms through linking phenomena occurring at the microscopic (e.g. pores and aggregates) to the mesoscopic (e.g. horizons and pedons) and the macroscopic (e.g. catenas and watersheds) levels, and eventually to the megascopic (e.g. regional and global) scales.

Hydropedology distinguishes itself from vadose zone hydrology (or soil hydrology) in the following aspects:

- Hydropedology emphasizes *in situ* soils in the landscape, where distinct pedogenic features (e.g. aggregation, horizonation, and redox features) and soil–landscape relationships (e.g. catena, soil distribution patterns, HFUs) are essential in understanding interactive pedologic and hydrologic processes, hence requiring adequate attention to soil architecture, soil layering, soil mapping, soil morphology, soil geomorphology, and stratigraphy.
- Hydropedology deals with the variably saturated and unsaturated zone in the surface and near-surface environments, including shallow root zone, deep vadose zone, temporally saturated zone, capillary fringe, wetlands, and even subaqueous soils (i.e. soils that form in sediment found in shallow permanently flooded environments such as in an estuary).
- Hydropedology is not just about soils’ impacts on hydrology; equally important is hydrologic feedbacks to pedogenesis, why subsurface heterogeneity exists, and how that impacts soil variability and soil functions.

As illustrated in Figure 1, hydropedology attempts to connect the pedon and landscape paradigms through linking phenomena occurring at microscopic (e.g. pores and aggregates) to mesoscopic (e.g. horizons and pedons) and macroscopic (e.g. catenae and watersheds) levels, and eventually to megascopic (e.g. regional and global) scales. Hydropedology, combined with hydrogeology, promotes an integrated systems approach to study the interactions of water with solid earth (i.e. soils, rocks, and anything in between) beneath the earth’s surface. Above the ground, hydropedology interfaces with hydroecology and hydrometeorology to understand the feedback mechanisms between climate,
vegetation, and soils. In the temporal dimension, hydropedology deals with both short- and long-term changes of the landscape–soil–hydrology relationships, including the use of pedogenesis and soil micro- and macromorphology as a signature of soil change and soil hydrology in the past and/or current conditions.

Hierarchical Frameworks for Bridging Forms and Functions in Hydropedology

Two hierarchical frameworks were suggested to bridge the forms and functions in hydropedology (Lin and Rathbun, 2003): one is soil mapping hierarchy that deals with soil spatial heterogeneity, and the other is soil modeling hierarchy that addresses temporal dynamics (Figure 2). The soil mapping hierarchy relates to the “forms” of soils that depict the spatial pattern of soil types or specific soil properties over the landscape of varying sizes, while the soil modeling hierarchy relates to the “functions” of soils that portray soil’s physical, chemical, and biological processes at different scales. Explicit links between these two hierarchies, however, have not yet been established. This is in part due to the mismatch of scales used in mapping and modeling. In a spatial sense, “aggregation” and “disaggregation” are used for mapping soil distribution, which are defined irrespective of a process-based model (Figure 2a). In contrast, “upscaling” and “downscaling” are used in modeling soil processes and/or quantifying model inputs/outputs, and thus are defined in the context of a specific model (Figure 2b). The meaning of “scale” also carries different meanings between mapping and modeling: in cartography, map scale refers to a ratio of map to reality, and the scale becomes smaller as spatial information is aggregated for a larger area; whereas in the modeling arena, scale is often used in a colloquial sense (without a specific quantifier), so large scale loosely refers to a large area extent.

Another constraint in connecting the forms and functions of soils for hydrologic modeling is the gap between scales used in either mapping or modeling hierarchies. Although there are approximations that might be taken...
to connect the scales of attribute-based mapping (such as spatial interpolation from point observations to areal coverages) or the scales of process-based modeling (such as extrapolation of smaller-area model parameters to larger-area values), significant knowledge gaps remain, and direct translation from one scale to another is still lacking. Since different factors or processes may dominate at different scales, and emerging new phenomena may occur as scale changes, a hierarchical set of models – based on available mapped attributes – may be constructed. As we focus on system patterns and filter out unimportant details, we might also begin to see emergent features that could serve as a natural skeleton to connect descriptions of soil and hydrologic responses across scales (Sivapalan, 2003). This could then form the basis of models that are inherently simpler at the macroscales, but with sufficient links to essential detailed heterogeneity and complexity observed at the microscales (McDonnell et al., 2007). Such an idea is somewhat similar to a hierarchy of identifiable dynamical patterns of processes, called successive dynamical levels, as suggested by Baveye and Boast (1999).

Jenny (1941) suggested the union of soil maps and soil functions to provide the most effective means of pedological research. At least two aspects hydropedology can contribute uniquely to such needed bridging of “forms” and “functions” and of different scales: One is through defining and quantifying soil architecture at various scales, and the other is through defining and identifying HFUs in different landscapes. Both efforts can advance hydrologic modeling via appropriate causal relationships and soil–landscape datasets of comparable resolutions.

**Defining a Broad Concept of “Soil Architecture” and its Quantification at Different Scales**

**Ped** (a naturally formed unit of soil aggregate such as a block, column, granule, plate, or prism) is a unique term in soil science. **Pedology** (a branch of soil science that integrates and quantifies the morphology, formation, distribution, and classification of soils as natural or anthropogenically modified landscape entities) captures that uniqueness well. The natural soil architecture is of essence in understanding soil physical, chemical, and biological processes as well as rainfall–runoff processes. It has been suggested that most that can be learned from sieved and repacked soil samples has already been done, so whatever we do with soil physics/hydrology should be done on intact, undisturbed soils, and preferably in situ in the field. A new era of soils research will have to rely on “structure-focused” – passing the stage of “texture-focused” – to achieve better ways of quantifying flow pathways, residence times, and spatiotemporal patterns of landscape water.

The term soil structure has been used in the US Soil Surveys, and elsewhere, to refer to the natural organization of soil particles into individual units (peds) separated by planes of weakness (Figure 3). Three attributes – ped grade, shape, and size – are used together to describe a soil structure. For example, “strong fine granular soil structure” means a soil that separates almost entirely into discrete units of peds (i.e. strong grade) that are loosely packed, roughly spherical (i.e. granular), and mostly between 1 and 2 mm in diameter (i.e. fine size). Some soils have simple pedality, each unit being an entity without component smaller unit; others have compound pedality, in which large units are composed of smaller units separated by persistent planes of weakness (Figure 3c).

Besides the shape, size, and grade of peds, the internal surface features of peds are also described in soil surveys, consisting of (i) coats of a variety of substances unlike the adjacent soil material and covering part or all of the surfaces, (ii) concentration of material on surfaces caused by the removal of other materials, and (iii) stress formations in which thin layers at the surfaces have undergone reorientation or packing by stress or shear (Soil Survey Division Staff, 1993). The kinds of structural surface features include clay films, clay bridges, sand or silt coats, other coats, stress surfaces, and slickensides (see examples in Figure 3). All of them have significant impacts on flow and transport processes in field soils.

Pores are considered separately in the U.S. Soil Survey’s concept of soil structure; however, in Europe, Canada, Australia, and some other countries, pore-related features (e.g. pore-size distribution, connectivity, and tortuosity) are an integral part of soil structure (e.g. Brewer, 1976). Thus, the US concept of soil structure is better termed pedality. Pedality and soil pore space are interrelated. But many soils have interpedal, intrapedal, and/or transpedal pores that are not necessarily represented by pedality. These pores, formed by biological activities (e.g. root channels and worm borrows), physical processes (e.g. desiccation cracking and freezing–thawing), or chemical reactions (e.g. dissolution or binding of soluble chemicals and organic matter) are critical in determining flow and transport in field soils. Therefore, in hydropedology, the term soil structure is used to encompass both pedality and pore space (Lin et al., 2005a).

Since soil structure generally refers to a specific soil horizon, a broader term of soil architecture is suggested here to also include the overall organization of a soil profile (e.g. horizonation), a soil’s relationship with the landscape (e.g. catena), and the overall hierarchical levels of soil structural complexity (Lin et al., 2005a).

Soil horizonation or layering is ubiquitous in nature, so it must be adequately addressed when measuring, modeling, and interpreting hydrologic processes in watersheds. Various kinds and thicknesses of soil horizons and how they organize in soil profiles reflect long-time pedogenesis and the past and current landscape processes. The fact that natural soils are layered has at least two significant
implications for hydrology: (i) interface between soil layers of contrasting textures and/or structures would slowdown water downward movement, which often leads to some kind of preferential flow (e.g. fingering flow, macropore flow, or funnel flow) and (ii) soil layering or discontinuity in soil hydraulic properties between layers would promote lateral flow or perched water table (PWT), especially in sloping landscapes with a water-restricting layer underneath.

A *catena* (also called *toposequence*) is a chain of related soil profiles along a hillslope with about the same age, similar parent material, and similar climatic condition, but differs primarily in relief that leads to differences in drainage and soil thickness. Catenary soil development often occurs in response to the way water runs down the hillslope and recognizes the interrelationship between soil and geomorphic processes (Hall and Olson, 1991; Moore *et al.*, 1993; Thompson *et al.*, 1997; Lin *et al.*, 2005a, b). Catenae are thus also often called *hydrosequences of related soils*, especially in depositional landscapes. Another important aspect of soil architecture along the hillslope is related to preferential flow network, which is further discussed later in this article.

While the importance of soil architecture across scales has long been recognized in soil science and hydrology, its quantification and incorporation into models have notoriously lagged behind. This problem is due to many factors, including those highlighted below:

- Inconsistent and fragmented concepts of soil structure, as described above and further elaborated by Letey (1991), Bronick and Lal (2005), Baveye *et al.* (2006), and many others. One critical point is illustrated here: "No art historian would attempt to characterize the architecture of buildings erected in France in the high Middle Ages by cataloging the nature of the stones used, the size of the stones, the types of materials used to bind the stones together, or the geometry of the portions of walls into which the buildings break when they are knocked down. With these parameters alone, it might be somewhat difficult to distinguish Notre Dame’s cathedral in Paris from stone houses in the Alps, for example. Clearly, a different perspective is needed, emphasizing the openings (window frames, doorways, rooms, etc.) that the stones allow to create in one case..."
or another. In the same manner, a focus on voids in soils might be a more fruitful way to look at soil architecture” (Letey, 1991; Young et al., 2001; Baveye et al., 2006).

- Overemphasis on ground-sieved soil materials and soil texture in the past, thus ignoring or downplaying the importance of the soil’s “natural architecture.” It has been said that a crushed or pulverized sample of the soil is related to the soil formed by nature in the same way as a pile of debris is related to a demolished building (Kubiëna, 1938). Lin (2007) also suggested that a crushed sample of the soil is as akin to a natural soil profile as a bulk of ground beef is to a live cow. The fundamental difference between \textit{in situ} soils and disturbed soil materials lies in soil architecture. In fact, soil is a living entity, with many dynamic forces acting upon it, so its internal architecture forms and evolves to serve various soil functions. That natural soils are structured to various degrees at different scales is the rule; whereas the existence of a macroscopic homogeneity is the exception (Vogel and Roth, 2003).

- Lack of appropriate techniques and devices to quantify soil architecture directly, especially \textit{in situ} noninvasively. Traditionally, soil structure has been evaluated by pedologists in the field using morphological descriptions or thin section observations, while soil physicists have employed wet and dry sieving, elutriation, and sedimentation to conduct aggregate analysis. In the absence of direct quantification, soil structure has been frequently evaluated by methods that correlate it to the properties or processes of interest (such as water retention, saturated hydraulic conductivity or $K_{sat}$, infiltration rate, and gas diffusion rate). In recent years, noninvasive methods that permit soils to be investigated without undue disturbance of their natural architecture have become increasingly attractive, which allow 3-D visualization of internal soil structure and its interactions with water. These methods include X-ray computing tomography, soft X-ray, nuclear magnetic resonance, $\gamma$-ray tomography, ground penetrating radar, and others (e.g. Anderson and Hopmans, 1994; Perret et al., 1999). Image analysis has brought new opportunities for analyzing soil structure, especially that of the pores, their sizes, shapes, connectivity, and tortuosity (e.g. Vogel et al., 2002; Vervoort and Cattle, 2003). However, although numerous attempts have been made to find either statistical relations or deterministic links between soil structural data and hydraulic properties, a significant gap remains between \textit{in situ} soil structure/architecture and field-measured soil hydraulic properties at different scales.

- Lack of a comprehensive theory of soil structure/architecture formation, evolution, quantification, and modeling that can bridge orders of magnitude in scale and integrate physical, chemical, biological, and anthropogenic impacts. Although a hierarchical organization of soil aggregates has been well recognized (Tisdall and Oades, 1982; Vogel and Roth, 2003) and fractal characterization of soil structure has been proposed (Perrier et al., 1999; Bartoli et al., 1998), there is still a lack of means of representing field soil structure in different scales in a manner that can be coupled into models of flow, transport, and rate processes (Lin, 2003; Lin et al., 2005a).

Quantifying soil architecture in the field across scales and at desirable spatial and temporal resolutions has been technologically limited. While landforms (e.g. digital elevation models or DEMs) and vegetation (e.g. land use/land cover) can now be mapped with high resolution (e.g. using Light detection and ranging or LIDAR and IKONOS earth observation satellite, respectively), there is a “bottleneck” phenomenon for \textit{in situ} high-resolution (e.g. sub-meter to centimeter) and spatiotemporally continuous and noninvasive mapping or imaging of subsurface architecture including flow networks. This technological bottleneck has constrained our predictive capacity of many soil and hydrologic functions.

\textbf{Defining “Hydropedologic Functional Unit (HFU)” and its Quantification in Different Landscapes}

At the hillslope scale, in the absence of detailed field investigations, much of soil information needed for hydrologic modeling is obtained from soil survey reports. However, the minimum size of soil map unit delineation in a standard soil survey is $\sim 1.6$ ha (Buol et al., 2003), which is often too coarse for most hillslope-scale studies. Consequently, many soil map units are associations and complexes, where different soil types are mapped together as a matter of necessity dictated by the map scale (Soil Survey Division Staff, 1993). Furthermore, polygon-based mapping in traditional soil surveys has resulted in soil variation being portrayed spatially as a step function that appears only when crossing from one polygon to another (Zhu et al., 2001). This does not allow the continuum of variability in soil attributes on a hillslope or within a catchment to be represented as it should. Future development of digital soil maps with greater spatial accuracy and resolution is needed to provide a better interface with other digital databases and to facilitate spatially distributed hydrologic modeling.

Traditional 3-D block diagrams used in soil surveys to show soil–landscape relationships can be enhanced to develop conceptual models of the landscape–soil–hydrology relationships (Figure 4). Enhanced 3-D block diagrams with added information of water table dynamics, water flow paths, hydric soils, restrictive layers, and other relevant information could significantly increase the values of soil survey products. As illustrated in Figure 4, classical block diagrams of soil–landscape relationships
could be used to indicate landscape hydrology, illustrating water flow direction and water-table dynamics. These block diagrams could be further linked to watersheds or physiographic regions to provide valuable conceptual frameworks of water movement over the landscape in different regions (USDA-NRCS, 1997).

New ways of mapping soils in greater detail and with higher precision are desirable for precision research and management. The concept of HFU, defined as a soil–landscape unit having similar pedologic and hydrologic functions, has been suggested as a means of cartographically representing important landscape–soil–hydrology functions (Lin et al., 2008). The goal of the HFU is to subdivide the landscape into similarly functioning hydropedologic units by grouping various geomorphic units that have similar storage, flux, pathway, and/or residence time of water in various soil–landscape units. These units can be identified and delineated using traditional soil survey methods and data in conjunction with various digital data sources (e.g. DEM), geophysical investigations, and in situ monitoring (Lin et al., 2008). Essentially, soil map units are better considered as landscape units rather than individual soil types (Wysocki et al., 2000). It is also desirable to go beyond the pure description of soil taxonomic classifications on a soil map (as traditional soil surveys did); rather, it would serve diverse applications much better by depicting soil functions, particularly those hydrologic functions critical to various local land uses. Hall and Olson (1991) have challenged traditional soil mapping: “Much effort has been expended on taxonomic classification of soils during the last few years but the importance of proper representation of landscape relations within and between soil mapping
units has been virtually ignored. The same mapping unit is often delineated on convex, concave and linear slopes. This mapping results in the inclusion of areas of moisture accumulation, moisture depletion and uniform moisture flow within a given mapping unit.”

Zhu et al. (2008) has proposed an approach toward precision soil mapping through integrating the existing second-order digital soil map with high-resolution terrain indices and site-specific geophysical surveys. Their refined soil map obtained for a 19-ha agricultural landscape had a considerably reduced variability within mapping units and an increased precision in soil boundaries. Such a map provided a basis for specific soil property mapping through addition of soil coring, or monitoring, or other means of soil sensing/mapping. On the basis of two soil properties important to hydrologic functions – surface soil texture and depth to clay layer, Lin et al. (2008) developed a map of HFUs for this agricultural landscape and tested this map for soil moisture dynamics and nitrogen fertilizer management. Their results showed that the three basic components making up an HFU (i.e. soil series or its variant, surface texture, and depth to clay layer) displayed significant differences in soil moisture. They also showed that an optimal combination of hydrology (both soil moisture storage and drainage) and nutrient supply (nitrogen fertilizer application rate) was required to obtain optimal growth of corn.

The HFU of different landscapes is a useful concept to be further developed. Such a mapping of HFUs can serve as a cartographic building block to increase the knowledge transfer and the bridging of “forms” and “functions” of soils at different scales. This is consistent with the flexible box models suggested by McDonnell (2003) and would facilitate the prediction of ungauged basins (Sivapalan, 2003).

There is a conceptual bottleneck that needs to be resolved for developing a new generation of hydrologic models. That is, should a continuous field or discrete objects be used to model surface and subsurface flow? Traditionally, hydrologic processes are generally conceptualized within the field domain (e.g. the Navier–Stokes equation and the Darcy’s law) (Goodchild et al., 2007). Classical hydrology has applied findings from fluid mechanics, together with the necessary constitutive relations, to develop sets of governing equations (much the same as atmospheric and ocean sciences have done). However, heterogeneities in land surface, hierarchical structures of soils, channel geometries, and preferential flow networks all make the land surface and subsurface different from the continuous-field assumption (Consortium of Universities for the Advancement of Hydrologic Science, Inc. (CUAHSI), 2007). It is becoming more and more recognized that solid earth is not a continuous fluid; rather, it poses hierarchical heterogeneities with discrete flow networks embedded in both the surface and the subsurface. As McDonnell et al. (2007), Kirchner (2006), Beven (2002), and many others have noted, current models in watershed hydrology are based on well-known small-scale physics or theories such as the Darcy’s law and the Richards equation built into coupled mass balance equations. It has been observed that the dominant process governing unsaturated flow in soils may change from matrix flow to preferential flow under different conditions when moving from the pore scale to the pedon scale (Blöschl and Sivapalan, 1995; Hendrickx and Flury, 2001). When moving from the pedon scale to the landscape/watershed scale, our knowledge for extrapolating the Darcy–Buckingham’s law and the Richards equation to large heterogeneous area is further constrained (Weiler and Naef, 2003). Further discussion on subsurface flow network is provided later in this article.

LINKING SURFACE/SUBSURFACE RUNOFF PROCESSES TO PEDOLOGIC UNDERSTANDING

In a new vision for watershed hydrology, McDonnell et al. (2007) urged a paradigm shift from merely gauging the net hydrograph at the outlet of a catchment (or at the trench of a hillslope) and using small-scale theories to model watershed hydrology to a new one that emphasizes diagnosis (i.e. diagnosis of soil–landscape patterns and watershed functional traits), classification (i.e. watershed classification based on similarity indices or dominant processes), evolution (i.e. why watershed heterogeneity exists), and flow network (i.e. underlying optimality principles) across scales. We believe pedology has significant potential to contribute to this new vision for watershed hydrology and to help answer “why-type” questions, as mentioned in the Introduction. In the following, we review various aspects of diagnostic soil–landscape features that are important to understanding and modeling surface/subsurface runoff processes at the three hierarchical scales of microscopic, mesoscopic, and macroscopic levels.

Aggregate and Macropore Scale

Pedality

Through aggregation, pedality alters a soil’s pore-size distribution and pore continuity/connectivity. Notably, the formation of peds creates interpedal pores that have size and shape different from intrapedal pores (Figure 3). Dye-tracing studies have clearly demonstrated that preferential flow often occurs along interpedal pores in structured soils (e.g. van Stiphout et al., 1987; Lin et al., 1996). Conceptually, we would expect the following trend of increasing capacity to transmit water vertically among different shapes of peds: platy < blocky, prismatic < granular (Figure 3a). Higher $K_{sat}$ has been reported for prismatic soil than its
blocky counterpart in silty clay loam (Bouma and Anderson, 1977a, b). Lin et al. (1999a) showed that, depending on texture and initial moisture, soils with prismatic peds had either higher or lower steady-state infiltration rates than similar soils with blocky peds. For fine-textured soils (clay and clay loam), prismatic pedality had higher flow rates than blocky pedality because of common occurrence of interprism macropores; whereas in medium-textured soils (sandy clay loam and silt loam), interprism pores were less significant in soils with prismatic pedality, leading to a lower water flow rate than in blocky soils (Lin et al., 1999a). The hydraulic distinctions between blocky and prismatic ped shapes can therefore be addressed through the associated difference in macroporosity and texture.

Conceptual understanding of pedality’s impacts on soil hydrology is intuitive, including flow pathway and flux (Figure 3). However, quantification of such impacts is less straightforward, in part because of the lack of a proper means for quantifying pedality. Pedality is usually expressed strongest at the surface and decreases with depth, resulting in a general trend of increasing ped size and decreasing ped grade with soil depth. We would also expect that the smaller the ped size, the higher the amount of interpedal pores, and that the stronger the ped grade (i.e. the distinctness or strength of ped units), the more significant the interpedal pores. However, tillage and other human and biological activities could significantly alter this general trend. Changing soil moisture also changes the expression of soil structure (especially in shrink–swell soils) and the relative volumes of peds and pores, adding to the complexity of finding mathematical solutions for modeling water movement in structured soils.

Spherical and blocklike pedality is most commonly seen in surface soils, while subsoils generally display blocklike or prismatic pedality. Platelet pedality is associated with surface sealing, compacted soil layer (such as plowpan), or sedimentation-inherited C horizon. Two structureless cases – massive (commonly associated with heavy clayey soils) or single grains (associated with sands) – are also observed in soils. Often, a soil horizon’s pedality is a mixture of several ped sizes (especially with compound pedality), with one dominant ped shape and grade. However, field qualitative descriptions of pedality often contain some degree of subjectiveness, leading to possible inconsistency among individuals who describe it.

In soils with compound pedality, Lin et al. (1999a) found that primary pedality had limited impact on infiltration at water potential of $\theta > -0.03$ m. Quisenberry et al. (1993) and Nortcliff et al. (1994) also reported that in a soil with both primary and secondary peds, preferential flow occurred mostly along secondary ped faces.

**Macropores**

Soil macropores encompass many types of structural openings in soils, such as desiccation cracks, pore space in between peds, animal burrows, and channels formed by living and dead roots. From a hydrologic functional point of view, these features often lead to preferential flow or bypass flow (Figure 3b). The definition of macropores, especially their size range, however, remains diverse. Some researchers have defined the size range of macropores on the basis of flow function: that is, whether flow is subjected to capillary force or not. Others have used morphological approach to define macropore size. It appears that both functional and morphological approaches should be combined to define macropore so that meaningful interpretations can be made in a comparable manner. As suggested by the Soil Science Society of America (SSSA, 1997), the minimal size of macropores is 0.075 mm in diameter. Macropores are also commonly defined as the pores corresponding to $\geq -0.03$-m water potential, that is, $\geq 0.5$ mm in radius for cylindrical pores or $\geq 0.5$ mm in width for planar pores. Most channels formed by the main roots of annual crops and earthworms have a minimal radius of about 0.5 mm. However, soil pipes formed by tree roots or animal burrows are often much larger, with pore diameter often greater than several centimeters.

Among various types of macropores, the general trend of increasing capacity to transmit water is (Lin et al., 1999a): vugh (isolated large void, usually irregular and not normally interconnected with other voids of comparable size) < channel (tubular-shaped void) < fracture (planar void between aggregates) < packing-void (void formed by random packing of single skeletal grains or peds that do not accommodate each other). Such a trend is supported by dye-tracing studies of Bouma et al. (1977) and Lin et al. (1996); both found the similar order as evidenced by dye-stained areas in soils with these types of macropores.

On the basis of steady infiltration rates measured on 96 soil horizons in Texas, Lin et al. (1999b) demonstrated that soil structure was crucial in characterizing hydraulic properties in macropore flow region, whereas texture had major impact on those hydraulic properties controlled by micropores. Lin et al. (1997) showed that steady-state infiltration rates at zero tension ($i_0$) were greatest for clay-textured soils with well-developed structure and macroporosity, while clayey soils with poorly developed structure had low infiltration rates, leading to high variability in $i_0$ for clayey soils. At more negative supply water potential, where macropores were less active, infiltration rates were greatest for soils dominated by sand.

**Soil Hydromorphology and Hydric Soil Field Indicators**

Actions of water on soils often result in the formation of distinctive morphological features that are indicative of field water movement. Soil horizons, for instance, often develop in response to water movement (such as leaching or accumulation of certain materials). Even without the
formation of visible morphological features, soil chemistry and biology may provide clues regarding field-scale water movement. For example, difference in mobility between redox-sensitive Mn and Fe in acid soil systems allows secondary Mn/Fe ratios to be used as podochemical indicators of field-scale throughflow (McDaniel et al., 1992) (Figure 5). A subset of soil morphological characteristics, known as hydric soil indicators (USDA-NRCS, 1998), is directly related to a specific set of hydrologic conditions. Soil morphology is also the testimony of long-term persistent flow and transport processes that result in visible pedological features such as clay films, tonguing, plinthites, and diverse soil structures that are of hydrologic significance (Figure 3). Soil macro- and micromorphology thus have long been used to infer soil moisture regimes, hydraulic properties, and landscape hydrologic processes (e.g. Bouma, 1992; Lilly and Lin, 2004; Lin et al., 2005a). Soil hydromorphy deals with soil morphological features (especially redoximorphic features or redox) caused by water and their relations with soil hydrology. Redox features (formerly called mottles and low chroma colors) are formed by the processes of alternating reduction—oxidation due to saturation—desaturation and the subsequent translocation or precipitation of Fe and Mn compounds in the soil (Soil Survey Staff, 1999). Types of redox features include (i) redox concentrations as accumulations of Fe/Mn oxides (e.g. nodules, concretions, masses, and pore linings), (ii) redox depletions as low-chroma (<2) features formed by removal of Fe oxides (including Fe depletions and clay depletions), (iii) reduced matrix that changes color upon exposure to air due to Fe(II) oxidation to Fe(III), and (iv) a reaction to an α,α-dipyridyl solution if the soil has no visible redox features (Vepraskas, 1992; Soil Survey Staff, 1999). Redox features, which are usually considered hydric soil indicators, exclude those hydric soil indicators composed of C and S (USDA-NRCS, 1998).

Hydric soils are defined as soils that formed under conditions of saturation, flooding, or ponding that lasted long enough during the growing season (repeated periods of more than a few days) to develop anaerobic conditions in the upper part (usually 0.15–0.3 m) of soil profiles (USDA-NRCS, 1998). Hydric soils are one of the three requirements (along with hydrophytic vegetation and wetland hydrology) for identifying jurisdictional wetlands in the United States. Hydric soil field indicators include a variety of features that are regional and texture-based, but all are formed predominantly by the accumulation or loss of Fe, Mn, C, or S compounds (USDA-NRCS, 1998). While indicators related to Fe/Mn concentrations or depletions are the most common, other features (e.g. sulfide and various combinations of carbon accumulations) have been used in specific kinds of soils that do not develop redox features. There are also so-called problem soils that are seemingly hydric soils but whose morphologies are difficult to interpret or seem inconsistent with the current landscape, vegetation, or hydrology (Veneman et al., 1998). These include soils formed in grayish or reddish colored parent materials, soils with high pH or low organic matter, Mollisols with thick dark A horizon, Vertisols with shrink–swell, soils with relict redox features, and disturbed soils such as cultivated soils and filled areas (USDA-NRCS, 1998; Rabenhorst et al., 1998). Relict redox features do not reflect contemporary or recent hydrologic conditions of saturation and anaerobiosis; rather, they were likely formed during past geological wetter climates. Typically, contemporary and recent hydric soil morphologies have diffuse boundaries, while relict redox features have abrupt boundaries (USDA-NRCS, 1998).

Certain redox patterns occur as a function of the patterns in which the ion-carrying water moves through the soil and as a function of the location of aerated zones in the soil. Characteristic color patterns are thus created by the reduced Fe and Mn ions removed from a soil if vertical or lateral water flow occurs, or the oxidized Fe and Mn precipitated in a soil if lack of sufficient water flux. Consequently, the spatial relationships of redox depletions and redox concentrations may be used to interpret water and air movement in soils (Vepraskas, 1992). Interpreting directions of water movement from redox patterns is easiest when the features have a consistent relationship with soil structure including macropores. For example, when redox depletions occur around macropores and redox concentrations occur within matrix, it would suggest that water infiltration has occurred along macropores and reducing condition has developed there because of perched saturated layers (Vepraskas, 1992). As another example, when redox

**Figure 5** Landscape distribution of dithionite-extractable Mn/Fe ratio (Mn$_{d}$/Fe$_{d}$) in the Bt soil horizon in a 3-ha field in the Piedmont region of North Carolina. The map was generated by kriging 60 sampling points occupying a variety of geomorphic positions. Dash lines indicate the center of topographic low areas.
concentrations occur around macropores and redox depletions occur within matrix, it would indicate that soil matrix is wet for periods long enough for reducing condition to be maintained while macropores become aerated because of faster drainage or plant roots (such as in rice plant) that transport air to macropores when the soil is still flooded (Vepraskas, 1992).

**Infiltration-excess Runoff**

A number of processes can cause the reduction in the soil’s infiltration capacity, leading to infiltration-excess runoff. These include surface sealing, soil freezing, hydrophobicity, and others. Infiltration-excess runoff events, however, can be minimized by proper management of the soil surface such as mulching and crop residue cover. The following elaborates each of these aspects.

- A surface seal or crust is caused by the breakdown of soil aggregates through raindrop impact or through a slaking process under saturated conditions (Ellison and Slater, 1945; Tackett and Pearson, 1965). Fine soil particles are dislodged from soil aggregates and are deposited as a thin layer over the soil surface. Through this process, large soil pores which would have been available for infiltration are “sealed” with the fine soil particles because the slaked clay particles are gel-like. Surface sealing and crusting occurs in virtually all arable soils during ponded infiltration or flood irrigation. The abrupt contact of the soil surface with excess water causes weak aggregates to disintegrate and slake. The migrating smaller particles can quickly form a seal on the soil surface within a short period of time.

- The infiltration capacity of a soil can also be greatly reduced when the soil freezes (seasonally or permanently). Concrete frost is often nearly impermeable and can be found most often in fine-textured soils (Zuzel et al., 1982; Dunne and Black, 1971). In agricultural areas, erosion from a partially thawed soil surface can be excessive. However, the infiltration capacity through soils that are initially dry or have large macropores is often minimally affected by soil freezing (Harris, 1972). The thick litter layer and large macropores found in typical forest soils minimize the potential effects of soil freezing on infiltration (Lutz and Chandler, 1946). Often in mountainous areas the deep snow cover insulates the ground and prevents soil freezing. Crawford and Legget (1957) observed that the frost depth was reduced about 0.3 m for every 0.3 m of undisturbed snow cover.

- Extensive field studies have shown that many soils are susceptible to hydrophobicity under dry conditions (Dekker and Ritsema, 2003). Many forest soils can become seasonally hydrophobic through leaching of the hydrocarbons in the litter layer and subsequent drying on the mineral soil; however, these soils become hydrophilic when soil moisture levels exceed 12–25% (Huffman et al., 2001). A hydrophobic soil layer is often developed after intense wild fires. Burning organic matter in a hot fire releases hydrocarbons into the atmosphere as a gas, which penetrates into soil aggregates at the mineral soil surface, condenses, and covers the aggregates with waxy substance after it cools (DeBano et al., 1970). This waxy coating repels water and greatly reduces the infiltration of the soil. Often, erosion from steep forested hillslopes after a hot-burning fire can be excessive. The thickness and persistence of hydrophobic layers varies primarily with burn severity; however, vegetation type, soil texture, and soil moisture are also important. Thick hydrophobic layers (i.e. >0.03 m) often persist for more than a year (Huffman et al., 2001).

- Additional complicating factors may also reduce a soil’s infiltration capacity, including soil swelling, soil instability, entrapped air, and soil layering. For example, infiltration into layered soil is always slowed down at the layer interface because of the hydraulic discontinuity (either different hydraulic conductivity or moisture holding capacity at a given soil water potential). In the case of a fine-textured layer overlaying a coarse-textured layer, “fingers” may develop after infiltrating water pore pressure overcomes the hydraulic discontinuity, forming fingering flow that may enhance the subsequent infiltration. Given the right condition, infiltration into structured, or macroporous, or cracking soil takes place in only a fraction of total soil surface, called fractional wetting (i.e. the fraction of total surface area that is wetted), thus vitiating the classical assumption of uniform advancement of the wetting front. Such fractional wetting may either reduce or accelerate infiltration depending on the initial and boundary conditions.

**Horizon and Profile Scale**

**Soil Profile Architecture and Water-restricting Soil Layers**

Soil profile is a vertical section of the soil through all its horizons and extending into the C horizon, while soil solum refers to only the portion of a soil profile above the C horizon. All materials above fresh, unweathered bedrock are also referred to as regolith, which is generally equivalent to a soil profile.

In the US Soil Taxonomy, a total of 8 diagnostic surface horizons (called epipedons) and 23 diagnostic subsurface horizons (called endopedons) have been identified, along with 30 additional diagnostic features for mineral soils (Soil Survey Staff, 2006). The presence or absence of these horizons plays the major role in determining which class a soil falls in Soil Taxonomy.

Numerous water-restricting subsurface soil horizons and features have been identified in Soil Taxonomy, including...
the following, with their major distinguishing characteristics, highlighted (genetic symbols are also shown in parenthesis for diagnostic horizons) (Soil Survey Staff, 2006):

- Duripan (Bqm): hardpan, strongly cemented by silica.
- Fragipan (Bx): dense pan, seemingly cemented when dry, brittle when moist, and slake when submerged in water. Typically has redox features and bleached prism ped faces.
- Glacial layer (Bf or Cf): massive ice or ground ice in the form of ice lenses or wedges.
- Ortstein (Bhm): cemented by humus and aluminum.
- Permafrost: a perennially frozen soil horizon that remains below 0 °C for ≥2 years in succession.
- Petrocalcic (Ckm): carbonate cemented horizon.
- Petrogypsic (Cym): gypsum cemented horizon.
- Placic (Csm): thin pan cemented by iron (or iron and manganese) and organic matter.
- Anhydrous conditions: a soil layer of cold deserts and other areas with permafrost (often dry permafrost) and low precipitation.
- Aquic conditions: continuous or periodic saturation and reduction, with redox features.
- Cryoturbation: frost churning on permafrost table, oriented rock fragments, and silt caps on rock fragments.
- Dense contact: a contact between soil and dense materials, without cracks or with ≥10 cm crack spacing.
- Dense materials: relatively unaltered but noncemented materials that roots cannot enter except in cracks (including till, volcanic mudflows, mine spoils, or other mechanically compacted materials).
- Fragic soil properties: essential properties of a fragipan, but lacks layer thickness and volume requirement for a fragipan.
- Gelic materials: materials that show cryoturbation and/or ice segregation in the active layer (seasonal thaw layer) and/or upper part of the permafrost.
- Lamellae: thin layer of oriented silicate clay accumulation.
- Lithic contact: boundary between soil and a coherent (often indurated) underlying material.
- Lithologic discontinuities: stone lines and others.
- Paralithic contact: contact between soil and paralithic materials, without cracks or with ≥10-cm crack spacing.
- Paralithic materials: partially weathered bedrock or weakly consolidated bedrock (such as sandstone, siltstone, or shale).
- Petroferric contact: boundary between soil and a thin ironstone sheet.
- Plinthite: dark red redox concentrations that changes irreversibly to ironstone on exposure to repeated wetting and drying.

Other subsoil horizons might also act as an aquitard or aquiclude to downward moving water, ultimately resulting in seasonal PWT and water moving laterally within the soil as subsurface throughflow (Busacca, 1989; Kemp et al., 1998). These include the following diagnostic subsurface soil horizons (with their genetic symbols indicated in parenthesis):

- Agric (A or B): silt, clay, and humus accumulation under cultivation just below plow layer.
- Argillic (Bt): silicate clay accumulation.
- Glossic (E): tongue resulting from the degradation of an argillic, kandic, or natic horizon.
- Kandic (Bt): argillic, kaolinite clays (low activity).
- Natric (Btn): argillic, high in sodium, columnar, or prismatic structure.
- Oxic (Bo): highly weathered, primarily mixture of Fe, Al oxides and nonsticky-type silicate clays.
- Spodic (Bh, Bs): organic matter, Fe and Al oxides accumulation.

In addition, stratified or dense geological materials (C or R horizons) also often develop a hydrologically restrictive layer that leads to PWT and lateral water movement. The importance of soil–bedrock interface and bedrock topography to subsurface stormflow has been recently recognized (e.g. Onda et al., 2001; Freer et al., 2002; McGlynn et al., 2002).

**Fragipan as an Example: Its Extent, Formation, and Impacts on Perched Water Table (PWT) and Hillslope/Catchment Runoff**

A query into the US Department of Agriculture (USDA) national soils database indicated that there are ~16.5 million ha fragipan soils in the United States, mostly distributed in the east and southeast (Figure 6). Other countries, such as New Zealand, Scotland, and Italy, have also reported fragipan soils (Soil Survey Staff, 1999). This very slowly permeable pan is one of the many water-restricting soil layers that frequently cause PWTs and control VSA hydrology. Soils having PWTs occupy ~82 million ha in the United States (McDaniel et al., 2008).

The processes that produce fragipans are imperfectly known, variously attributed to physical ripening, the weight of glaciers, permafrost processes, and other events during the Pleistocene (Smeck and Ciolkosz, 1989; Soil Survey Staff, 1999). Some of the properties of fragipans are inherited from buried paleosols. Fragipans most commonly occur in soils that formed under forest vegetation. The upper boundary of most fragipans has a narrow depth range of about 0.5–1 m below the surface, if the soil is not eroded. It is interesting to note that fragipans form only in soils in which water moves downward through the profile and they are commonly at depths that rarely freeze (Soil Survey Staff, 1999). The polygonal network of bleached materials in fragipans is formed by reduction
of free iron after water has saturated the cracks (Figure 6 inset).

Dense, slowly permeable fragipans are important controlling factors in VSA hydrologic processes in heterogeneous watersheds (Gburek et al., 2006). Needelman et al. (2004) showed that hillslopes with fragipan-containing soils produced considerably more runoff than did those without fragipans. Even within a relatively homogeneous catchment comprising one fragipan-containing soil series, the variable depth to the fragipan also gives rise to VSA hydrology, which in turn influences the spatial pattern of runoff generation (Figure 7). Many years’ monitoring in the Palouse Region of northern Idaho and eastern Washington by McDaniel et al. (2008) showed that the spatial distribution of saturation-excess runoff is largely controlled by the distribution of fragipan and its varying depth from the surface (Figure 7). Areas of the catchment where fragipans are relatively close to the surface tend to become saturated more quickly and generate more saturation-excess runoff. Subsurface lateral flow on top of the fragipan accounts for redistribution of up to 90% of the precipitation received on site during late winter and early spring in the study by McDaniel et al. (2008). The combination of low drainable porosity, relatively shallow depth to the fragipan, and relatively high lateral $K_{sat}$ in surface soil layers create a very “flashy” hydrologic system (McDaniel et al., 2008).

**Saturation-excess Runoff**

Strictly speaking, the term saturation means 100% pore space filled with water. In reality, field soil saturation is more a “satiation,” meaning <100% pore space filled with water (because of air bubbles etc.), but with free water present. Three types of soil saturation/satiation are defined in soil survey (Figure 8) (Soil Survey Staff, 2006):

- Endosaturation: The soil is saturated with water in all layers from the upper boundary of saturation to a depth of 2 m or more from the mineral soil surface.
- Episaturation: The soil is saturated with water in one or more layers within 2 m of the mineral soil surface and also has one or more unsaturated layers with an upper boundary above a depth of 2 m, below the saturated layer(s). This is commonly associated with PWT on top of a relatively impermeable layer. The gleying by episaturation is sometimes called pseudo-gley or false gley, in contrast to true gley caused by endosaturation.
- Anthric saturation: This is a special kind of saturation that occurs in cultivated and irrigated (flood irrigation) soils.
Depending on the duration of saturation and whether or not reduction occurs, two conditions are associated with soil saturation (Soil Survey Staff, 2006):

- Aquic conditions: continuous or periodic saturation and reduction occur, leading to redox features (gleying).
- Oxyaquic conditions: saturated but not reduced and thus no redox features. This may be related to transient water table or flashy soil hydrologic conditions.

A combination of topography, soil architecture below the surface, and vegetation largely determines the location and frequency of saturation-excess runoff. Key soil attributes critical to understanding the processes leading to saturation-excess runoff include (i) depth to a hydrologically restrictive soil or bedrock layer, (ii) change in $K_{sat}$ and drainable porosity with soil depth, (iii) change in bulk density and saturated moisture content with soil depth, (iv) soil horizonation and effective vertical versus lateral $K_{sat}$, and (v) preferential flow in soil profiles. Each of these factors is further discussed in the following:

- The depth to a hydrologically restrictive layer is one of the most critical parameters in understanding VSA hydrology. This depth, which could be either the distance from the soil surface to a soil horizon that limits vertical water movement or a bedrock, is often referred to as the soil depth in VSA hydrology. Obviously, for the same amount of water input, a shallow soil will reach saturation leading to runoff sooner and more frequently than a deeper soil. The soil depth not only determines the maximum water volume that a soil profile can hold before runoff is generated, but also
directly relates to the water volume that a soil profile will hold before a PWT is developed. At this point, large macropores become active leading to downslope lateral flow or subsurface stormflow (Whipkey, 1965). Once this subsurface stormflow is generated, water follows the topography of this impeding layer. Recent studies have provided clear evidence that water often follows bedrock topography rather than the soil surface topography as assumed in many hydrologic models (Burns et al., 1998; Freer et al., 2002). Although soil depth and the topography of water-impeding layer are clearly important factors for understanding hillslope and catchment hydrology, there are few tools available to reliably map these features over large areas.

- Soil $K_{sat}$ and drainable porosity are commonly assumed to have exponential decline with soil depth in hydrologic models (Beven and Kirkby, 1979; McDonnell, 2003; Weiler et al., 2005). This holds in many columns from organic-rich top layers to more compacted restrictive subsoil layers (Kendall et al., 1999; Brooks et al., 2004, 2007). For example, Beven (1982) demonstrated using experimental data taken from hillslope hydrology experiments on 24 soils in predominantly forested landscapes that both $K_{sat}$ and soil porosity decreased exponentially with depth below the soil surface. These exponential decay functions, however, were not as consistent when later examining 38 soils from the broader database presented by Holtan et al. (1968), (Beven, 1984). When deeper soil profiles are considered, that is, including C horizons and soil layers beneath water-restrictive layers (Figure 8), then such an exponential decay function is misleading. As illustrated by West et al. (2008) for soils in the Southern Piedmont of Georgia, five of ten sites had a lower soil horizon’s $K_{sat}$ being higher than the overlying horizons. When all sites lumped together, mean $K_{sat}$ in C horizons was 2.5 times higher than the overlying BC and Bt2 horizons (West et al., 2008). Lin (1995) also documented that field $K_{sat}$ (approximated by steady-state infiltration rates at zero tension) in many of the 18 soil profiles measured in Texas did not follow exponential decline trend from the soil surface to $\sim 2$ m depth. Drainable porosity in deep, freely draining soils is often defined as the difference between saturated moisture content and field capacity. In soils dominated by saturation-excess runoff due to a restrictive layer, however, free drainage does not occur as it would in deep soils. During the wet season a PWT forms, which on sloping ground drains primarily by lateral flow or has very little drainage on relatively flat ground. At equilibrium soil water above the water table is held under pressures equivalent to the elevation above the water table. Thus, the drainable porosity for shallow soils is the difference between saturated moisture content and the moisture content at the soil surface, which

Figure 8  An illustration of the difference between episaturation and endosaturation: episaturation is from the top and endosaturation is from the side or bottom. This cross section of a landscape also shows the regional versus perched water tables in relation to three soils, one well drained and two with poor internal drainage. The soil containing the perched water table is wet in the upper part, but unsaturated below the impermeable layer, and therefore is said to be epiaquic, while the soil saturated by the regional water table is said to be endoaquic. This cross section also illustrates the importance of subsurface architecture for understanding hillslope hydrology (From Brady and Weil, 2004).
Saturated soil moisture (which can be derived from bulk density) has a similar effect on the generation of surface runoff as soil depth, because it determines soil water storage capacity. Soils having low saturated moisture content can store less overall water, causing more frequent and greater amounts of saturation-excess runoff. Like \( K_{\text{sat}} \), saturated moisture content typically decreases with soil depth from organic-rich top layers to compacted restrictive layers, resulting in less overall water storage at greater soil depths (Beven, 1982, 1984; Brooks et al., 2007). The bulk density of the soil is also the most widely used parameter to estimate the amount of open pore space. Pedotransfer functions have used bulk density along with soil texture to incorporate variability in soil structure into effective \( K_{\text{sat}} \) measurements. However, these parameters by themselves are often more appropriate for explaining micropore flow rates rather than macropore flow as macropores are often not well represented by small soil cores used to determine bulk density (Beven, 1984; Lin et al., 1999b).

Soil horizonation impacts the overall effective vertical versus lateral \( K_{\text{sat}} \) in a soil profile. The depth distribution of vertical \( K_{\text{sat}} \) is important in defining where in the soil profile will begin to saturate (i.e. defining the depth to a restrictive layer) and the relative magnitude of lateral \( K_{\text{sat}} \) will determine runoff patterns in a catchment dominated by VSA hydrology. Surface runoff will be confined primarily to toe slopes or at sharp breaks in the slope from steep to flat in soils having large lateral \( K_{\text{sat}} \). Surface runoff patterns will be more widespread and correlated with variability in soil depth in areas with low lateral \( K_{\text{sat}} \) (Dunne et al., 1975). The challenge for hydrologists trying to predict subsurface lateral flow or stormflow at the hillslope scale is to determine representative lateral \( K_{\text{sat}} \). Using small soil cores it is relatively easy to measure \( K_{\text{sat}} \) in the laboratory but these measurements may not represent the \( K_{\text{sat}} \) at larger scales. Without a method to physically characterize soil architecture at the hillslope scale, especially large macropores and their network, \( K_{\text{sat}} \) becomes a calibration parameter in most hydrologic models (Grayson et al., 1992; Wigmosta et al., 1994). Most likely, the calibrated \( K_{\text{sat}} \) values will not be able to adequately represent the flow of water through a hillslope without accounting for differences in soil structure in soil profiles and how macropores are distributed within each soil layer. Tracer and isotope studies often reveal that simplified representations of how water flows through a hillslope are inadequate to represent the true flow path (Vaché and McDonnell, 2006). A hillslope-scale experiment conducted by Brooks et al. (2004) showed that the hillslope-scale \( K_{\text{sat}} \) was 15, 5, and 2 times greater than the small-scale measurements in the A, Bw, and E soil horizons, respectively.

### Preferential Flow in Soil Profiles

Preferential flow is the process whereby water and dissolved chemicals move by preferred pathways at an accelerated speed through a fraction of a porous medium. Vervoort et al. (1999) suggested that preferential flow in soil profiles is related to soil structural differences (e.g. macropore flow and fractional flow) or textural differences (e.g. fingering flow and funnel flow). Different pedological features are indicative of possible preferential flow in various field soils, especially if combined with landscape observations. These include: (i) soil structural features (such as pedality, coatings, macropores, and slickensides); (ii) redox features (indicating water-restricting layer and nonuniform water movement); (iii) water-restrictive soil layers (such as sloping lamellae indicating the likely occurrence of funnel flow); and (iv) lithologic discontinuities (indicating significant changes in particle-size distribution and thus possible fingering flow or other types of preferential flow). A simple field technique was devised by Bouma (1997) to measure preferential flow in clay soils as a function of rain intensity/quantity and initial soil moisture content. Deriving cracking patterns from theoretical soil swell–shrink characteristics turned out to be difficult, and very small pores (such as slickenside fissures shown in Figure 3) with a volume that cannot be measured with existing physical methods can conduct large volumes of water (Lin et al., 1996).

Flühler et al. (1996) have suggested three regimes of flow and transport within a soil profile during a preferential flow process: (i) lateral distribution flow in the attractor zone where preferential flow is initiated; (ii) downward preferential flow in the transmission zone where water moves along preferential flow pathways and bypasses a considerable portion of the soil matrix; and (iii) lateral and downward dispersive flow in the dispersion zone where preferential flow pathways are interrupted and water flow becomes more or less uniform again. This model appears to capture the general characteristics of vertical preferential flow in many soil profiles, and is also consistent with the depth function of soil hydraulic conductivity discussed earlier.

On the basis of a review of studies documenting soil microbial biomass distributions with soil depth (e.g. Van Gestel et al., 1992; Dodds et al., 1996; Richter and Markewitz, 2001; Blume et al., 2002; Taylor et al., 2002; Fierer et al., 2003), the system of Flühler et al. (1996) can be applied. Previous studies have demonstrated a consistent...
pattern of changing microbial biomass size and activity distribution, characterized by three broad zones with distinct biogeochemical potentials that correspond to the three flow zones of Flühler et al. (1996): (i) attractor zone, equivalent to the A horizon and representing high-biomass surface soils under the strong influence of plant roots, moisture, and temperature; (ii) transmission zone, equivalent to B and C horizons and representing lower more spatially heterogeneous biomass than in the attractor zone (Konopka and Turco, 1991; Rodríguez-Cruz et al., 2006); and (iii) dispersion zone, representing higher-biomass, water-capture zones at the soil–bedrock interface or right above a water-restricted soil layer because of moisture and more nutrient-rich conditions resulting from impeded or lateral flow (Buss et al., 2006). These three zones are distinguished from “deep-subsurface” aquifers and geologic materials, many of which have extremely low biomass (Kieft et al., 1995).

**Catena and Catchment Scale**

**Topography: Linking Hydrology with Pedology**

The topography of a landscape is responsible for providing the overall direction of water flow and hydraulic gradient for downslope subsurface stormflow (Dunne et al., 1975; see also Chapter 112, Subsurface Stormflow, Volume 3). In VSA hydrology, the spatial variability of soil moisture before an event and of subsurface stormflow during the event dictates where saturation-excess runoff will occur on the landscape. Topographically convergent toe slopes often are the wettest locations leading to saturation-excess runoff.

In addition to largely controlling the distribution and magnitude of water flow through a landscape, topography is also considered one of the five major factors controlling soil formation. In the classic work by Russian geologist V.V. Dokuchaev and later extended by Jenny (1941), the five major soil-forming factors include climate, organisms, topography, parent material, and time. Topographic patterns dictate the way in which mass and energy are distributed within a landscape. The orientation of a hillslope also dictates the amount of solar radiation received, and therefore largely controls the rates of evaporation, transpiration, and snow ablation. The topography controls wind fields leading to variability in evaporation, snow drift, and deposition. The topography also greatly influences rates of soil erosion and deposition. The combination of these effects leads to spatial patterns of vegetation diversity, biomass production, and soil variability (see Chapter 12, Co-evolution of Climate, Soil and Vegetation, Volume 1).

There is a close feedback between hydrology and soil development within a landscape that is driving by differences in topography. Topography drives the distribution and transport of water and sediment, which in turn influences soil development (e.g. the development of restrictive layers through eluviation/illuviation processes). This soil development can then feedback to the variability in water distribution and transport over the landscape.

With the ever-increasing numeric processing power available today and the availability of spatially detailed elevation maps, hydrologists have been able to better represent landscape processes using various means from simple topographic indices to detailed simulation models. Geographic information system (GIS) models can use detailed elevation maps to describe topographic influences on water fluxes throughout a landscape. For example, Moore et al. (1991) summarized a wide range of topographic attributes that can be used to map various hydrologic, geomorphic, and biological processes. The most well-known static topographic index for representing the influence of topography on the spatial distribution of saturation-excess runoff that utilizes the upslope catchment area and local slope was incorporated into TOPMODEL (Beven and Kirkby, 1979). This simple wetness index utilizing only topography has been successful in describing areas generating saturation-excess runoff patterns during wet seasons. However, growing evidence has pointed out the critical importance of the topography of underlying bedrock or water-restricting soil layer in defining water flow paths and hence soil wetness distribution (e.g. Onda et al., 2001; Freer et al., 2002; McGlynn et al., 2002; McDonnell, 2003).

**Catena**

Spatial distribution of topographic attributes that characterize water flow also capture spatial variability of soil attributes at the hillslope scale. Because of the processes that occur along a hillslope, soils may be quite different in different portions of a landscape, but these processes and relationships may be similar across a larger area, particularly if geomorphology and stratigraphy are similar. Numerous investigations of catenae have been completed in order to address this phenomenon (e.g. Veneman and Bodine, 1982; Evans and Franzmeier, 1986; Pennock and de Jong, 1990; Stolt et al., 1993; Khan and Fenton, 1994; Thompson et al., 1998). It is now believed that the greatest variability in certain soil properties within a physiographic region may occur along a hillslope rather than from one side of the region to the other (e.g. Pennock and de Jong, 1990; Lin et al., 2005b). This warrants a careful investigation and understanding of hillslope soil variability and catenae before making broad generalizations about soil variability within a physiographic region.

Water table depth (drainage condition) and fluxes of water, solutes, and sediments typically differ in soils along a catena (Schaetzl and Anderson, 2005). However, catenae in different climatic and physiographic regions may exhibit markedly different relationships between soil and hydrologic properties. In many low-relief landscapes of humid regions, proximity of a water table to the soil surface increases with distance away from stream or drainage way.
An example of this can be seen on the broad, flat, low-relief Atlantic coastal plain in the southeastern United States. The most poorly drained soils are found toward the centers of the broad interstream divides, while the most well-drained soils are restricted to the edges of the flats and slopes closest to the streams and estuaries (Daniels et al., 1971; Daniels et al., 1984). The opposite trend is seen in the higher-relief landscapes associated with the nearby Piedmont region. Water table proximity to the soil surface decreases with increasing distance away from the streams or drainage ways. As a result, the well-drained soils occupy the uplands and the more poorly drained soils occupy the lower slope positions near the streams (Daniels et al., 1984). Similarly, fluxes associated with overland flow, subsurface lateral flow, percolation, capillary rise, and return flow will also vary along a catena in different climatic and physiographic regions (Schaetzl and Anderson, 2005; Sommer and Schlichting, 1997). These fluxes will vary spatially and temporally, resulting in VSA hydrologic processes.

Two opposite trends of soil depth and soil wetness distribution along steeply sloped catenae are illustrated here. In the humid forested catchment developed from shale on the ridge of central Pennsylvania (within the Ridge and Valley physiographic region), soil thickness and wetness generally increase in concave hillslopes from hilltop to valley floor, while that of the convex and planar hillslopes within the same catchment shows no clear difference from the hilltop to the hill bottom (Lin et al., 2006b). In contrast, in the rugged Hill Country of central Texas with stair-step topography developed from limestone (part of the Edwards Plateau physiographic region), soil thickness and infiltration capacity decrease from the hilltop (upper riser with steeper slopes) to the hill bottom (tread with flatter slope) (Wilcox et al., 2007). Wilcox et al. (2007) also found that the upper riser subsoils were saturated or very wet for extended periods, and surface runoff was generated from lower part of the hillslopes (lower risers and treads) during periods of high rainfall. Therefore, runoff and erosion increased from upper riser to tread. Despite the opposite trends, both catenae described above demonstrate the controls of soil, topography, and geology on runoff generation (Lin et al., 2006b; Wilcox et al., 2007).

**Subsurface Preferential Flow Networks in Hillslopes and Catchments**

Uhlenbrook (2006) claimed that catchment hydrology is a science in which all processes are preferential. It has been well recognized that preferential flow can occur in just about any soil, leading to spatial concentrations of water and chemicals that are not well described by the classical approach based on uniform flow assumption (Vervoort et al., 1999; Hendricks and Flury, 2001). A large body of research on preferential flow and on hillslope hydrology has been accumulated in the past decades (e.g. Kirby, 1978; Beven and Germann, 1982; Gish and Shirzolmammad, 1991; Germann and Hensel, 2006). However, because of the complex and dynamic nature of preferential flow in different geographic regions and at different space- and timescales, our ability to predict landscape variability in water flow and solute transport has been limited (Jury, 1999; Lin et al., 2005a, 2006a). This, in part, is due to the fragmented knowledge of quantitative relationships between preferential flow and soil structure, soil layering, landscape features, precipitation inputs, and antecedent moisture conditions at different scales. It is interesting to note that most preferential flow studies by the soil science community has focused on the pore to pedon scales and vertical flow, while the hydrology community seems to have focused more on the hillslope to catchment scales and lateral flow. A more concerted and integrated effort will advance our predictive capacity in truly 3-D flow system.

There is an emerging recognition of a paradigm shift from a “continuum”-based approach to a “network”-based view of hillslope and watershed hydrology (Rinaldo et al., 2006; Uhlenbrook, 2006; Sidle et al., 2001; McDonnell et al., 2007). A “3F” (form, function, and feedback) is needed to characterize various watershed networks, which would lead to a better understanding, modeling, and prediction of hillslope hydrology and ungauged basins. It remains to be further tested, whether, though an internal network structure exists in the subsurface of many hillslopes, which appears to govern vertical and lateral preferential flow dynamics and the threshold-like hydrologic response under varying precipitation, soil, and antecedent moisture conditions (e.g. Weiler et al., 2005; Tromp-van Meerveld and McDonnell, 2006; Lin and Zhou, 2008). These issues are important because without a clear understanding of the initiation, connectivity, and persistence of preferential flow networks across space and time, we are unable to effectively predict rainfall–runoff for the right reasons.

Analogous to stream branching networks, a hillslope/landscape network is a set of linkages among different reservoirs and pathways with varying conductance of flow. During storms with wet soil conditions, such a network often provides preferred pathways for water to flow downgradient with high velocities (Nimmo, 2007). Threshold phenomena may occur at critical “nodes,” or junctions of a network (i.e. hot spots), where significant changes may occur within a short period of time (i.e. hot moments). Threshold behavior of subsurface storm-flow has been observed in many hillslopes and landscapes (Whipkey, 1965; Mosley, 1979; Peters et al., 1995; Weiler et al., 2005; Tromp-van Meerveld and McDonnell, 2006; Lin and Zhou, 2008). Thus, the concept of subsurface preferential flow networks could provide a strong scientific advance in our understanding of seemingly complex
Interaction of macropore with the surrounding or adjacent soil matrix resulting in expanded preferential flow path

Bedrock
Mineral soil
Organic-rich horizon

Wet
Wet

Macropore flow
Uneven bedrock topography

Preferential flow through living and decayed roots

Formation of a perched water table temporarily between organic-rich and mineral soil horizons

Heterogeneity caused by preferential flow pathways

Impeding layer but some water flow into cracks and fractures

Water movement through humus and loose soil zones

Formation of a perched water table above bedrock

Water emerging from fractured bedrock into soil

Some macropores connecting to the bedrock fractures

Preferential flow

Matrix flow
Downslope direction
Vertical direction
Resultant direction

Formation of a perched water table temporary between organic-rich and mineral soil horizons

Highly permeable relative to the mineral soil and very high density of roots

Hydrologic (and biogeochemical) phenomena across landscapes. Sidle et al. (2001); Gish et al. (2005); Lin (2006), and many others have reported evidence of preferential flow self-organization in forested hillslopes and agricultural landscapes, where individual short preferential flow pathways are linked via a series of “nodes” in a network, which may be switched on or off, or expand or shrink depending on local soil moisture conditions and landscape locations (Figure 9). Sidle et al. (2001) suggested different levels of nodes in a network to approximate the preferential movement of water and solutes over the landscape, including (i) those that are easily activated under various conditions (including critical landscape locations such as topographic depressions, contrasting soil layers and soil–bedrock interfaces, and direct physical linkages between mesopores and macropores), (ii) those that require very wet conditions to activate (such as disconnected large macropores, buried pockets of organic matter, and discontinuous animal burrows and soil pipes), and (iii) those that do not promote preferential flow (such as soil matrix and dead-end macropores with no connecting nodes). This conceptual model could lead to a new way of modeling preferential flow networks across multiple scales.

SUMMARY AND FUTURE OUTLOOK

It is clear that synergies can be generated by bridging classical pedology with soil physics, hydrology, and geomorphology to advance the mechanistic understanding and process-based modeling of rainfall–runoff processes. The challenge of getting the right answer for the right reason in hydrologic modeling and prediction (Grayson et al., 1992; Kirchner, 2003; McDonnell et al., 2007) calls for contributions from hydropedology and other related disciplines to enhance the understanding of interactive pedologic and hydrologic processes and the complex landscape–soil–hydrology relationships across space and time. This article has illustrated that hydropedology can make unique contributions to advancing hillslope and watershed hydrology in many ways, among which are the following key areas that require further research:

- Better understanding and quantification of soil architecture from the pore to the catchment scales, and its relationships to preferential flow dynamics across space and time;
- Defining and mapping hydropedologic functional unit (HFU) in different landscapes and its incorporation into spatially distributed hydrologic modeling;
- Breakthroughs in two bottlenecks – a new conceptual/theoretical framework for modeling preferential flow networks (which is different from the continuous-field assumption) and a technological advancement in nondestructively mapping/imaging subsurface architecture;
- Pattern recognition and prediction of the “forms” and “functions” of soils and watersheds across scales, which could offer comprehensive insights regarding heterogeneity and the underlying processes leading to heterogeneity;
- Use of various tracers, including isotopes and microbes, in combination with advanced sensor networks, long-term environmental monitoring, and a new generation of computer models, to gain insights into optimality
principles underlying the watershed functional traits (McDonnell et al., 2007; also see Chapter 12, Co-evolution of Climate, Soil and Vegetation, Volume 1, Chapter 13, Pattern, Process and Function: Elements of a Unified Theory of Hydrology at the Catchment Scale, Volume 1).

Acknowledgments

The fragipan map of the United States (Figure 6) was generated with help from Sharon Waltman of the USDA-NRCS and Steven Crawford of Penn State University. The paragraph on the summary of soil microbial distributions with soil depth and its possible relationships with the three zones of preferential flow was assisted by Mary Ann Bruns of Penn State.

REFERENCES


