

# Hydrologic Discovery Through Physical Analysis: Honoring the Scientific Legacies of Wilfried H. Brutsaert and Jean-Yves Parlange

Abstracts from Special Section of Water Resources Research



Jean-Yves Parlange and Wilfried Brutsaert, May 2012 at the Symposium held May 14-15, 2012, Cornell University, Ithaca, New York, USA (Photo Robert Cooke)

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# Introduction: Learning from the scientific legacies of W. Brutsaert and J.-Y. Parlange

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## Abstract

Though the essence of the scientific literature is to be a repository of unaffiliated truths, scientific advancement fundamentally stems from the insights and efforts of individuals. This dichotomy can hide exemplars for young scholars of how to contribute to scientific understanding. This section of *Water Resources Research* addresses eminently successful strategies for advancement of the science of hydrology by exploring the ramifications of the work from Drs. Wilfried Brutsaert and Jean-Yves Parlange, colleagues who made many of the most significant contributions to the understanding of hydrologic processes of the last 50 years. The generous scope of the special section follows the key areas of their contributions, but the content looks forward from their work. Important and novel results span solute transport, infiltration, streamflow generation, and evaporation.

**Keywords:** solute transport, infiltration, streamflow generation, evaporation, groundwater, erosion

This issue traces key branches of the scientific legacy stemming from seminal contributions of two scientists who both recently retired from Engineering at Cornell University, Drs. Wilfried Brutsaert and Jean-Yves Parlange. The contributions included here make clear the explanatory power of the approach shared by Brutsaert and Parlange: the application of concise physically-based representations of processes to understand myriad and intertwined elements of the water cycle. Underlying these works is the recognition that the most effective way to provide mechanistic understanding, while being guided by Ockham's razor, is to base analyses upon the governing equations for the formative processes. The parallel seen in their strategies extends also to the recognition these men received: both are members of the National Academy of Engineering, awardees of the AGU's Horton Medal and Horton Award, and AGU Fellows.

Articles for the special section were sought in two formats: **Reviews** (focusing on original works of the Brutsaert and Parlange and developing the contemporary lines of discovery that emerged from these roots), and **Letters**, which are short contributions that offer focused single-idea advancements on fields that were pioneered by Drs. Brutsaert and/or Parlange. The reviews include (working from the center of the earth outwards): DiCarlo's presentation on the current state of understanding of fingered flow in unsaturated media; Lo and Sposito's presentation of propagation of acoustic waves in porous media; Assouline's survey of quantitative models for infiltration into soils; Troch et al.'s exploration of understanding the temporal behavior of recession of river flow through the lens of the Boussinesq Equation; Barry et al.'s survey of contaminant and sediment transport literature which grew from Parlange's approaches; and Dias' compilation outgrowths from Brutsaert's contributions to models and concepts in the parameterization of atmospheric turbulence. The letters were

succinct articles which fit under the umbrella provided by the reviews, to render broad literature reviews redundant. Here we find original contributions inspired by work of Brutsaert and Parlange including: advancement of compact approaches to modeling infiltration (Triadis and Broadbridge); effects of capillary forces on water table fluctuations (Kong et al.); the validity of pan evaporation data to estimate landscape evapotranspiration (Brutsaert); and complementary relationships for estimation of evaporation (Crago and Qualls).

While Parlange and Brutsaert have long been close friends and colleagues, and collaborated in hydrological research, it would be more accurate to describe them as working in close parallel than together for the majority of their contributions. The many cross-citations of each other's results started in the 1970's, but it was not until 1987, after Parlange had moved to Cornell, that they co-authored their first joint paper. Seven papers joint papers followed. Undoubtedly, their mutual admiration and discussions together with Robert Miller enhanced their research endeavors and those of many others at Cornell University. Apart from their scientific insights, it is in this latter aspect, the direct and indirect effects on other researchers, which will be their most long-lived legacy. Their research records attest to the many cooperative research efforts they have undertaken, as suggested by long lists of co-authors. Second, their students and research collaborators are found in institutions worldwide, following lines of research derived from their interactions with Brutsaert and Parlange.

We sincerely appreciate the extraordinary effort put forward by the authors of the papers in this issue, and the tireless assistance of the AGU staff in the preparation of this special section. Most of all, we give our deepest gratitude to Drs. Parlange and Brutsaert for providing the guidance and daily inspiration, which has immeasurably enriched our lives and the vigor of the advancement of hydrologic sciences.

## Infiltration into soils: Conceptual approaches and solutions

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[1] Infiltration is a key process in aspects of hydrology, agricultural and civil engineering, irrigation design, and soil and water conservation. It is complex, depending on soil and rainfall properties and initial and boundary conditions within the flow domain. During the last century, a great deal of effort has been invested to understand the physics of infiltration and to develop quantitative predictors of infiltration dynamics. Jean-Yves Parlange and Wilfried Brutsaert have made seminal contributions, especially in the area of infiltration theory and related analytical solutions to the flow equations. This review retraces the landmark discoveries and the evolution of the conceptual approaches and the mathematical solutions applied to the problem of infiltration into porous media, highlighting the pivotal contributions of Parlange and Brutsaert. A historical retrospective of physical models of infiltration is followed by the presentation of mathematical methods leading to analytical solutions of the flow equations. This review then addresses the time compression approximation developed to estimate infiltration at the transition between preponding and postponding conditions. Finally, the effects of special conditions, such as the presence of air and heterogeneity in soil properties, on infiltration are considered.

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### 1. General Introduction

[2] When a wetting fluid enters in contact with dry porous medium, flow is induced in the porous material due to capillarity. When this movement occurs in the absence of gravity forces (as in horizontal flow, for example), it is considered as a “sorption” process. When it occurs under the influence of gravity, it is considered as an “infiltration” process. In the following review, we will limit ourselves to that specific case of wetting under gravity, and infiltration will be defined, following *Brutsaert* [2005], as “the entry of water into the soil surface and its subsequent vertical motion through the soil profile.”

[3] Infiltration is a complex process, which depends upon a large number of factors: water supply rate; the elapsed time since the onset of water application; soil and water chemical compositions; spatial variability and distribution of the hydraulic properties within the soil profile; initial and boundary conditions; topography; temperature; and probably additional factors linked to biological and microbiological activities in the soil. Wetting of a porous medium consists in a multitude of events where the wetting fluid invades fully or partially empty pores, crevices, or cavities. It is thus a phenomenon that involves a series of microscale

processes that affect its macroscale behavior. However, most of our conceptual and quantitative tools regarding the infiltration process rely on observations and measurements pertaining to that macroscale reflection of the events occurring at the pore scale. This introduces an intrinsic “tension” between the phenomenon in reality and the conceptual approaches and models developed during the years to describe it. This tension is exacerbated in some specific cases when the conditions induce flow instability, and the models reach their limits of applicability.

[4] Different types of approaches were developed to provide quantitative tools able to describe and predict infiltration into porous media in general, and soils in particular. Based on a large body of experimental studies that provided the community with valuable data from laboratory and field experiments, empirical expressions were first proposed. Then, mathematical solutions were derived from the basic physical model expressed in terms of differential equations. With the apparition of computers, numerical models were also developed that could solve directly these differential equations, thus releasing some of the assumptions and constraints needed to reach amenable mathematical solutions.

[5] Infiltration can accommodate the entire water supply rate or only part of it. The “infiltration capacity,” or the potential infiltration rate, of a soil [*Horton*, 1940] is the maximal rate at which the soil surface can absorb water. Therefore, when the water supply exceeds the infiltration capacity, only part of it infiltrates and the remaining part ponds on the soil surface or runs off according to local topography. The infiltration capacity function can be thus considered a soil characteristic with dependence on the initial soil water content profile. When infiltration is below its capacity, the infiltration function will also depend upon the temporal history of the application rate. Figure 1 illustrates

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## Solute and sediment transport at laboratory and field scale: Contributions of J.-Y. Parlange

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[1] We explore selected aspects of J.-Y. Parlange's contributions to hydrological transport of solutes and sediments, including both the laboratory and field scales. At the laboratory scale, he provided numerous approximations for solute transport accounting for effects of boundary conditions, linear and nonlinear reactions, and means to determine relevant parameters. Theory was extended to the field scale with, on the one hand, the effect of varying surface boundary conditions and, on the other, effects of soil structure heterogeneity. Soil erosion modeling, focusing on the Hairsine-Rose model, was considered in several papers. His main results, which provide highly usable approximations for grain-size class dependent sediment transport and deposition, are described. The connection between solute in the soil and that in overland flow was also investigated by Parlange. His theory on exchange of solutes between these two compartments, and subsequent movement, is presented. Both deterministic and stochastic approaches were considered, with application to microbial transport. Beyond contaminant transport, Parlange's fundamental contributions to the movement of solutes in hypersaline natural environments provided accurate predictions of vapor and liquid movement in desert, agricultural, and anthropogenic fresh-saline interfaces in porous media, providing the foundation for this area of research.

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### 1. Introduction

[2] Mass transport is a core factor in the analysis and prediction of environmental quality, for example, as a control on time scales of environmental system resilience. Apart from quantifying key elements of environmental system response, models of fate and transport are central to contaminant data analysis, risk assessment, and prognostic modeling,

to name but a few. Diffuse environmental pollution is ubiquitous [e.g., Carey *et al.* 2013; Islam and Tanaka, 2004; Novotny, 1999, 2007; Posen *et al.*, 2011]; thus techniques for environmental protection and remediation rely on the accuracy of models that predict outcomes of alternative strategies for remediation. It is no surprise then that modeling approaches are heavily embedded in analysis of transport processes. For example, a search within the 5500 papers published in *Water Resources Research* from 2000 to the present revealed that two-thirds include “model” in the title, abstract, or keywords. Nearly one in five papers includes both “model” and “transport” in these search categories.

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[3] J.-Y. Parlange has made a vast array of contributions to environmental mass transport. Here, we focus on solute and sediment transport leaving, for example, his extensive work on water flow to be described by Assouline [2013]. Later, we explore his contributions to mass transport in overland flow (including sediment transport and transfers to flow from the surface soil) and in the near subsurface. Additionally, we briefly examine his contributions to thermodynamics of soil solutions. Our objectives are, first, to provide a guide to his body of work in this domain and, second, to give a flavor of his approach, which is both theoretical and physically based. Table 1 is intended to satisfy the first objective. The second objective is addressed in the following sections.

### 2. Column-Scale Solute Transport

[4] Column-scale solute transport is described by the classical advection-dispersion equation (ADE) [e.g., Barry, 1990; Bear, 1972]:

## Late-time drainage from a sloping Boussinesq aquifer

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[1] Numerical solutions to the nonlinear Boussinesq equation, applied to a steeply sloping aquifer and assuming uniform hydraulic conductivity, indicate that late-time recession discharge decreases nearly linearly in time. When recession discharge is characterized by  $-dQ/dt = aQ^b$ , this is equivalent to constant  $dQ/dt$  or  $b = 0$ . This result suggests that a previously reported exponential decrease with time ( $b = 1$ ) of modeled recession discharge from a similar sloping aquifer represented by the same equation appears to be an artifact of the numerical solution scheme and its interpretation. Because the linearly decreasing recession discharge ( $b = 0$ ) is not known from field studies, these findings challenge the application of a nonlinear Boussinesq framework assuming uniform conductivity and geometric similarity to infer hydraulic properties of sloping aquifers from observations of streamflow. This finding also questions the validity of the physical interpretation of the exponential decline in late time resulting from the commonly used linearized form of the Boussinesq equation, opposed to the full nonlinear equation, when applied under these conditions. For this reason, application of the linearized equation to infer hydraulic properties of sloping aquifers is also challenged, even if the observed recession is consistent with that of the linearized Boussinesq equation.

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### 1. Introduction

[2] Following the pioneering recession flow analysis work by *Brutsaert and Nieber* [1977], *Troch et al.* [1993], and *Szilagyi et al.* [1998], attempts have been made to use analytical solutions to (linearizations of) the nonlinear Boussinesq equation to infer hydraulic properties of sloping aquifers, e.g., while using the linearized Boussinesq equation [*Huyck et al.*, 2005], focusing on the rising limb of a hydrograph [*Pauwels and Troch*, 2010], or testing a hill-slope subsurface flow similarity number [*Lyon and Troch*, 2007]. These attempts make use of the method of hydrograph analysis, first introduced by *Brutsaert and Nieber* [1977], when considering recession flow from horizontal aquifers. *Brutsaert and Nieber* [1977] showed how under certain initial and boundary conditions, the recession discharge  $Q(t)$  from a horizontal Boussinesq aquifer can take the form

$$-\frac{dQ}{dt} = aQ^b \quad (1)$$

where  $a$  is a function of the aquifer properties and  $b$  is a constant. For instance, for flat-lying aquifers, characterized by spatially and vertically uniform saturated hydraulic conductivity  $k$  and drainable porosity (or specific yield)  $f$ , it can be shown analytically [*Brutsaert and Nieber*, 1977] that after initial effects are dissipated (“late time”)

$$a = 4.8038 \frac{k^{1/2} L_d}{fA^{3/2}} \quad (2a)$$

$$b = 3/2 \quad (2b)$$

with  $A$  catchment area and  $L_d/A$  drainage density.

[3] The direct use of equation (2a) to infer, for instance,  $k$  from recession data and known  $A$ ,  $L_d$ , and  $f$ , requires that  $b = 3/2$  and that the assumptions underlying the nonlinear Boussinesq equation hold. Any  $b \neq 3/2$  thus raises challenges for the use of equation (2a) to infer  $k$  for flat-lying aquifers. Indeed, in applications of equation (2a), *Brutsaert and Nieber* [1977] and *Troch et al.* [1993] showed that their discharge data suggested  $b = 3/2$  during late time. *Eng and Brutsaert* [1999] and *Brutsaert and Lopez* [1998] found  $b = 1$ , and consequently replaced equation (2a) with a similar approach based on the linearized Boussinesq equation, that does correspond to  $b = 1$ . Using an alternative set of assumptions, especially a drawdown that is small relative to saturated thickness, *van de Giesen et al.* [2005] demonstrated that solutions to aquifer recession based on the Laplace equation leads to  $b = 1$  during late time as well.

[4] It bears noting that the use of a similar approach to infer aquifer properties for sloping hillslopes or catchments

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## Investigating storage-discharge relations in a lowland catchment using hydrograph fitting, recession analysis, and soil moisture data

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[1] The relation between storage and discharge is an essential characteristic of many rainfall-runoff models. The simple dynamical systems approach, in which a rainfall-runoff model is constructed from a single storage-discharge relation, has been successfully applied to humid catchments. Here we investigate (1) if and when the less humid lowland Hupsel Brook catchment also behaves like a simple dynamical system by hydrograph fitting, and (2) if system parameters can be inferred from streamflow recession rates or more directly from soil moisture storage observations. Only 39% of the fitted monthly hydrographs yielded Nash-Sutcliffe efficiencies above 0.5, from which we can conclude that the Hupsel Brook catchment does not always behave like a simple dynamical system. Model results were especially poor in summer, when evapotranspiration is high and the thick unsaturated zone attenuates the rainfall input. Using soil moisture data to obtain system parameters is not trivial, mainly because there is a discrepancy between local and catchment storage. Parameters obtained with direct storage-discharge fitting led to a strong underestimation of the response of runoff to rainfall, while recession analysis leads to an overestimation.

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### 1. Introduction

[2] In most catchments, discharge depends strongly on the total amount of water stored. Head differences cause groundwater to flow toward channels and quick runoff processes such as overland flow, subsurface stormflow, or drainpipe flow occur only when the catchment is sufficiently wet. Because the relation between storage and discharge is essential to describe runoff generating processes, it is a major component of most rainfall-runoff models: in physically based models, groundwater flow is computed from differences in groundwater levels between cells, conceptual models contain one or more reservoir components, and in the simplest models the whole catchment is represented as one (non)linear reservoir.

[3] It would be advantageous if the storage-discharge relation could be easily derived from a limited amount of observations or catchment characteristics, which would allow application in the desired model without calibration. This has been a hydrological research topic for many decades. *Brutsaert and Nieber* [1977] investigated discharge recession curves, *Kirchner* [2009] used system parameters obtained with recession analysis in a simple hydrological model, and *Teuling et al.* [2010] used this model to obtain

system parameters by means of calibration. There are many other studies in which storage-discharge relations have been found by means of discharge (recession) analysis, but a more obvious approach, namely using storage data directly, is not known to the authors. However, local storage computed with a groundwater model has been used in direct storage-discharge fitting [*Rupp et al.*, 2009]. Examples of studies in which solutions to the Boussinesq equation for sloping aquifers have been employed in order to investigate storage-discharge relations are *Troch et al.* [1993], *Brutsaert* [1994], and *Rupp and Selker* [2006b]. We refer to the review article by *Troch et al.* [2013] and references therein for a more complete overview.

[4] While storage-discharge relationships have been investigated in detail for mountainous catchments and in humid climates, it is unclear whether the approach can be extended to lowland catchments or less humid climates. Lowland catchments cover an extensive part of the world's most densely populated areas and, therefore, adequate discharge forecasts in these catchments are of large societal and economic value. *Kirchner* [2009] used the Plylimon catchment (runoff ratio = 0.79) to illustrate his simple dynamical systems approach, in which a nonlinear reservoir model is based on the storage-discharge relation. *Teuling et al.* [2010] applied this approach to the slightly less humid Rietholzbach catchment (runoff ratio = 0.73), where the approach worked well during wet conditions, but failed during dry summers.

[5] Here, we investigate (1) if and when the less humid lowland Hupsel Brook catchment (runoff ratio = 0.39) also behaves like a simple dynamical system by hydrograph fitting, and (2) if system parameters can be inferred from discharge recession rates or, more directly, from biweekly

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# Use of pan evaporation to estimate terrestrial evaporation trends: The case of the Tibetan Plateau

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[1] There is still no general agreement on the relationship between the evaporation of open water from a small pan and the terrestrial evaporation from the surrounding landscape under drying conditions. A possible way out of this impasse is reviewed and applied to the harsh and extreme climatic conditions of the Tibetan Plateau. It is confirmed herein that during 1966–2000 with a pan evaporation trend of  $-4.57 \text{ mm a}^{-2}$ , the terrestrial evaporation trend was  $+0.7 \text{ mm a}^{-2}$ , in agreement with the experimental findings of Zhang et al. (2007).

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## 1. Introduction

[2] Probably the most obvious and simplest way of studying natural evaporation is to measure the rate at which water vaporizes into the air from a pan, a dish, or some other similar open container. Well known among the earliest such studies are those conducted by Halley [1687, 1694] in London at Gresham College, Sedileau [1730, 1733] in 1692 in Paris at the Royal Observatory, Van Musschenbroek [1739] in Utrecht in his own garden, and Dalton [1802] in Manchester.

[3] Over the years, more formal and routine investigations were initiated to determine the water requirements for crop irrigation in arid areas and the evaporative water losses from the attendant reservoirs. Probably one of the oldest standardized pans is the Colorado Sunken Pan, going back to the work of Carpenter [1889, 1891] and Rohwer [1934]; it has a square water surface with sides of 91.5 cm (3 ft), a depth of 45.7 cm (1.5 ft), and its rim about 10 cm (4 in) above the surrounding ground surface. But the instrument now in widest use is the class A pan of the U.S. Weather Bureau [Kadel and Abbe, 1916]. It is a cylindrical container, with inside dimensions of a 121.9 cm (4 ft) diameter and a 25.4 cm (10 in) depth; it is made out of non-rusting metal and is placed 3–5 cm above the ground on an open-frame wooden platform. Also widely used is the GGI-3000 pan, which was developed in Russia [World Meteorological Organization, 2008]. It is buried in the ground as a cylindrical container down to 60 cm, with slightly conical bottom down to 68.7 cm at the center; its rim is at 7.5 cm above the ground surface. Of a different design is the 20 cm diameter pan, which has been used at more than 1736 stations in China [Liu and Kang, 2007] since at least the early 1950s [Yang and Yang, 2012, Figure 2]. It has a circu-

lar surface with a bird guard around it; its surface is placed at about 70 cm above the ground, and as it has a depth of 10 cm, the bottom of this pan is at 60 cm above the ground. It is now gradually being replaced by a bigger pan similar to the GGI-3000 [Fu et al., 2009; Xiong et al., 2012].

[4] Although originally not intended for this purpose, more recently attempts have been made to use pan evaporation measurements also in the estimation of evaporation from the landscape under natural, i.e., nonirrigated, conditions. But there is still no general agreement on how this can be accomplished. Accordingly, it is the present objective first, briefly to review a likely way out of this impasse and to clarify the issues at hand; and second, to apply the resulting approach to illustrate the recent evolution of terrestrial evaporation in one of the harsher and more extreme climatic environments on Earth, namely, the Tibetan Plateau. In what follows for simplicity's sake the term evaporation denotes the combined vaporization of water from vegetation and soil, and it has the same meaning as the alternative term evapotranspiration.

## 2. Pan Evaporation Versus Terrestrial Evaporation

### 2.1. Moist Surface Covered by Vegetation With Adequate Water Supply

[5] First observe that the terrestrial evaporation  $E$ , under conditions of adequate water supply, is in fact the potential evaporation  $E_{po}$ , as originally defined by Thornthwaite [1948], as follows:

When water supply increases, as in a desert irrigation project, evapotranspiration rises to a maximum that depends only on the climate. This we may call 'potential evapotranspiration,' as distinct from actual evapotranspiration.

[6] This concept of potential evaporation is somewhat ambiguous for a number of reasons [e.g., Brutsaert, 2005, p. 130]; consequently, some authors have preferred to use a "reference evapotranspiration" calculated on the basis of the Penman-Monteith approach to characterize the evaporation  $E_{ws}$  from the well-maintained lawn with adequate water supply at the weather station; unfortunately, because

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## An exact series and improved numerical and approximate solutions for the Boussinesq equation

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[1] We derive a series solution for the nonlinear Boussinesq equation in terms of the similarity variable of the Boltzmann transformation in a semi-infinite domain. The first few coefficients of the series have been known for a long time, having been obtained by a truncated inversion of the series solution of the Blasius equation, but no direct recurrence relation was known for the complete series representing the solution of the Boussinesq equation. The series turns out to have a finite radius of convergence, which we estimate with a numerical complex-plane integration method that identifies the singularities of the solution when the equation is extended to the complex plane. The homogeneous condition at the origin produces a singularity which complicates numerical solutions with Runge-Kutta methods. We present two variable transformations that circumvent the problem and that are best suited to the complex-variable and the real-variable versions of the equation, respectively. Using those tools, an approximate solution accurate to  $1.75 \times 10^{-10}$  and valid for the entire positive real axis is then developed by matching a Padé approximant of the exact series and an asymptotic solution (to overcome the restriction imposed by the finite radius of convergence of the series), along the same lines of the expression proposed by Hogarth and Parlange (1999). The accuracies of all of the existing and the newly proposed solutions are obtained.

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### 1. Introduction

[2] The Boussinesq equation for groundwater flow in an unconfined aquifer [Boussinesq, 1904] is a nonlinear partial differential equation and it has no known analytical solution for arbitrary boundary conditions. However, a few exact analytical solutions are known, each one for specific boundary conditions: the solution that Boussinesq himself developed for a finite domain [Boussinesq, 1904], the series solution derived by Barenblatt [1952] for a time-dependent water head at the origin (whose recurrence relation and proof of convergence was presented by Song *et al.* [2007]) and the solution derived by Parlange *et al.* [2000].

[3] Although the derivation of the Boussinesq equation involves some approximations (such as the Dupuit hypothesis and the negligibility of the nonsaturation effects), it is still an accurate mathematical model for some cases [see Steenhuis *et al.*, 1999; Verma and Brutsaert, 1971]. As a result of this, the Boussinesq equation remains a topic of

active research in hydrology. Recently, Moutsopoulos [2013] obtained approximate analytical and numerical solutions for it subjected to nonlinear Robin boundary conditions and Basha [2013] derived an approximate result using the traveling wave approach. It is also used to model the propagation of a gas in porous media, nonlinear mass diffusion in solids and heat diffusion in solids [Moutsopoulos, 2010].

[4] In this work, we solve the Boussinesq equation for a horizontal one-dimensional aquifer (as depicted in Figure 1) with no recharge, viz.

$$\frac{\partial h}{\partial t} = \frac{k_0}{n_e} \frac{\partial}{\partial x} \left( h \frac{\partial h}{\partial x} \right), \quad (1)$$

with initial and boundary conditions

$$h(x, 0) = H, \quad h(\infty, t) = H, \quad h(0, t) = 0, \quad (2)$$

where  $h$  is the water table height,  $t$  is time,  $k_0$  is the saturated hydraulic conductivity, and  $n_e$  is the drainable porosity.

[5] By means of the Boltzmann transformation

$$\phi(\xi) \equiv \frac{h(x, t)}{H}, \quad \xi \equiv \frac{x}{\sqrt{4Dt}}, \quad (3)$$

where  $D = Hk_0/n_e$ , equation (1) with initial and boundary conditions given by equation (2) can be reduced to

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## The value of intuitive concepts in evaporation research

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[1] The Evaporative Fraction (EF) and the Complementary Relationship (CR), both extensively explored by Wilfried Brutsaert during his productive career, have elucidated the conceptual understanding of evapotranspiration within hydrological science, despite a lack of rigorous proof of validity of either concept. We briefly review Brutsaert's role in the history of these concepts and discuss their appeal and interrelationship.

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### 1. Introduction

[2] Part of the legacy of Wilfried Brutsaert's hydrological research is the value he has placed on using intuitive concepts to understand hydrologic processes. This letter discusses two of the concepts he has employed in his evaporation research: the conservation of flux ratios during the daytime, and the CR between regional and potential evaporation. While neither has been rigorously proven, both have been surprisingly fruitful in promoting our understanding of evaporation processes, are widely used in research, and have been incorporated into operational models and data products (representative references will be given in the course of this letter).

[3] In this letter, we will: (1) Briefly review Brutsaert's role in the history of the conservation of flux ratios and the CR; (2) Discuss what about these concepts accounts for their appeal and fruitfulness; and (3) Discuss their interrelationship.

### 2. History

#### 2.1. Conservation of Flux Ratios

[4] The evaporative fraction  $EF = LE/(H + LE)$  is the ratio of latent heat flux  $LE$  to available energy  $H + LE (=R_n - G)$ , where  $H$  is sensible heat flux,  $R_n$  is net radiation, and  $G$  is ground heat flux. The value of  $EF$  indicates how favorable conditions are for conversion of available energy into latent rather than sensible heat fluxes [e.g., Brutsaert, 2005]. The concept of the “conservation,” “self-preservation,” or “self-similarity” (all terms used by Brutsaert himself) of  $EF$  during a single daytime is usually credited to Shuttleworth *et al.* [1989], using data from the FIFE experiment in Kansas in

1987. They observed that  $EF$  remained nearly constant on fair-weather days. They built on earlier work by Jackson *et al.* [1983] who took the ratio  $LE/R_s$ , where  $R_s$  is incoming shortwave radiation, as constant during the daytime [see Brutsaert, 2005]. Sugita and Brutsaert [1991] and Nichols and Cuenca [1993] also used constant  $EF$  to extrapolate from a single  $LE$  measurement to the total daytime evaporation. Brutsaert and Sugita [1992] extended the analysis to other “flux ratios” which they defined as the ratio of  $LE$  to “any other flux component,” finding that good estimates of daytime evaporation could be obtained assuming the self-preservation of  $EF$ ,  $LE/R_n$ , and  $LE/R_s$ . Brutsaert and Chen [1992] plotted  $EF$  during the daytime for a succession of days during a drying period of FIFE 1987. They observed conservation of daytime  $EF$  throughout the drying period, with lower values on successive days.

[5] Brutsaert also explored reasons for the self-preservation. Brutsaert and Sugita [1992] noted that clouds disrupt the diurnal progress of  $EF$ . Crago and Brutsaert [1996] showed, by propagating uncertainties through the equations, that uncertainty in daytime evaporation estimated by assuming constant  $EF$ , is generally smaller than with constant Bowen ratio  $H/LE$ .

[6] Crago [1996] searched systematically for a physical reason for the self-preservation of  $EF$ . Other studies looked to convective boundary layer models to analyze flux ratio self-preservation [e.g., Lhomme and Elguero, 1999; Peters-Lidard and Davis, 2000; Raupach, 2000; Gentine *et al.*, 2010, 2011a, 2011b]. Gentine *et al.* [2010, 2011a, 2011b] also considered temporal patterns of variability of the constituents of flux ratios. Van Niel *et al.* [2012] developed a general model to evaluate sources of error and to minimize bias within conservation of flux ratios [see also Van Niel *et al.*, 2011]. Thus, there is a considerable body of work providing a theoretical understanding of the preservation of flux ratios. The consensus is that some of the flux ratios, including  $EF$ , tend to have only small variability during the daytime under conditions of relatively large, stable net radiation, and steady, low-to-moderate horizontal advection [cf. Brutsaert, 2005, p. 138].

#### 2.2. Complementary Relationship

[7] The CR between actual regional-scale evaporation  $E$  and apparent potential evaporation  $E_p$  was first proposed by

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## Research on atmospheric turbulence by Wilfried Brutsaert and collaborators

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[1] In this paper, several lines of investigation of atmospheric turbulence, by Wilfried Brutsaert, his students, and collaborators, are reviewed. Overall, we classify these lines as *K*-theory, surface roughness parameterization dealing with momentum and scalar fluxes, radiative effects on temperature fluctuations, stable conditions, scalar similarity, and atmospheric boundary-layer parameterization. Emphasis is placed on turbulence parameterization. Although these topics are presented more or less in chronological order, this order is broken whenever connections need to be established. Hopefully, these connections are the most interesting part of this review: it is there that Brutsaert's insights and long-range scientific questions may be found. His approach invariably included a careful formulation of the physical and mathematical basis of the problem at hand, and proceeded to focus on some essential issues that allowed analytical, numerical or statistical treatment. There is much to be learned from this approach; it is hoped that some of it can be glimpsed here.

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### 1. Introduction

[2] The atmospheric boundary layer is the region of the atmosphere closest to the surface of the Earth and directly impacted by the surface fluxes of momentum, sensible heat, water vapor, and several gases of interest in air pollution and biogeochemical studies [Brutsaert, 1982; Stull, 1988; Garratt, 1994; Seinfeld and Pandis, 1998]. Therefore, the connection between the ABL features and the regional, or basin-wide, surface fluxes is evident: understanding the ABL is essential for identifying the relevant scales of most hydrological problems; to decide which processes to represent explicitly, and which to parameterize, and more generally for improving our hydrological models, as made clear by Brutsaert [1986].

[3] Most of the time, the ABL is turbulent, particularly in the case of the convective, daytime boundary layer, when the surface fluxes are largest. Then, prompted by the practical needs of hydrology, one is naturally drawn into studying ABL turbulence, a situation not unlike many other cases where engineering demands motivated the advancement of fluid mechanics. The archetypal flow is the turbulent boundary layer (originally studied for pipes and channels, see, e.g., Darrigol [2005]), which not only gives

the name to the ABL but is the starting point for many concepts that prove fruitful also for atmospheric turbulence.

[4] As it happens with many researchers, the approaches used by Wilfried Brutsaert and his collaborators to a better understanding of atmospheric turbulence that we will be reviewing here follow many of the early paths taken by fluid dynamicists to study turbulent flows. At the same time, however, they added substantial improvements as these approaches needed adaptation to the particularities of the ABL: its inherent nonstationarity, the surface geometry imposed by nature, the surface spatial inhomogeneities, the effects of buoyancy, and many other facets, all required a concerted effort to adapt, and sometimes create, the concepts needed to model ABL turbulence.

[5] The focus of the present work is atmospheric turbulence itself. Thus, in section 2, we review the results derived from semi-empirical, or *K*-theory, for the behavior of surfaces ranging from evaporating pans to natural lakes, as well as insights on the form of the turbulent diffusivity tensor itself. In section 3, we review contributions to the parameterization of scalar mass and heat-transfer coefficients with the identification of scalar roughnesses different from those for momentum. Section 4 gives a brief overview of results concerning the relative importance and direct dissipative effects of longwave radiation in ABL turbulence. Section 5 deals with results for the stable surface layer and section 6 with the important issue of how different the turbulent transport characteristics of two scalars are. In a sense, this is a development of the ideas started with the so-called Reynolds' analogy [Reynolds, 1900] to the study of turbulent transfer of different scalars. Section 7 then deals with results for the convective boundary layer and its interface with the top of the surface layer where local free convective conditions prevail. They are also related to drag, mass, and heat transfer

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## Stability of gravity-driven multiphase flow in porous media: 40 Years of advancements

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[1] Gravity-driven multiphase flow in porous media is ubiquitous in the geophysical world; the classic case in hydrology is vertical infiltration of precipitation into a soil. For homogenous porous media, infiltrations are sometimes observed to be stable and laterally uniform, but other times are observed to be unstable and produce preferential flow paths. Since Saffman and Taylor (1958), researchers have attempted to define criteria that determine instability. Saffman and Taylor's analysis consisted of two regions of single phase flow, while Parlange and Hill (1976) integrated this analysis with the multiphase flow equations to provide testable predictions. In the subsequent 40 years, great advances have been made determining the complex interactions between multiphase flow and instability. Theoretically, the stability of the standard multiphase flow equations has been verified, showing the necessity of extensions to the multiphase flow equations to describe the observed unstable flow. Experimentally, it has been shown that the instability is related to a phenomena in 1-D infiltrations called saturation or pressure overshoot. In this review, the connection between overshoot and instability is detailed, and it is described how models of overshoot can simplify the analysis of current and future models of instability and multiphase flow.

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### 1. Introduction

[2] In hydrology, the classic case of gravity-driven multiphase flow in porous media is infiltration of water in soil. Stability of the flow field is a key question for infiltration, as the formation of preferential flow paths can create large consequences on the transport of contaminants to ground and surface waters. The filtering capacity is lost when infiltration occurs through a few preferential pathways, effectively bypassing the soil. For example, during the 1999 NY state fair, hundreds of people were sickened and one toddler perished after a summer storm caused *E. coli* from animal feces to enter the supposedly safe ground water drinking supply through preferential flow paths [Yarze and Chase, 2000]. There are many implicated specific causes of preferential vertical flow of water, for example, natural heterogeneities in the soil (wormholes and roots) or soil layering [Gerke *et al.*, 2010; Steenhuis *et al.*, 1998], but the general cause is that the water is heavier than the gas that it is displacing.

[3] This review, and the associated papers in this issue (R. Wallach and Q. Wang, On using Miller similarity to scale sorptivity by contact angle, submitted to *Water Resources Research*, 2013), concentrates on gravity-driven preferential flow in uniform porous media, i.e., with no macroscopic heterogeneities [Glass *et al.*, 1991]. Here preferential flow occurs when the initially flat flow front becomes unstable and breaks up into many discrete flow paths, and thus is often referred to as a flow instability problem. In addition to ground water protection, correct models of gravity-driven flow instability are crucial to predicting subsurface flows such as transport of radionuclides [Bundt *et al.*, 2000], recovery of hydrocarbons through water and/or gas flooding [Lake, 1989], or injection of CO<sub>2</sub> into saline aquifers [Szulczewski *et al.*, 2009].

[4] Figure 1 shows cartoons of a water front infiltrating into a porous media. On the left, the infiltration is stable with a laterally uniform front where the small frontal perturbations do not grow. On the right, the infiltration is unstable, where the small perturbations have grown into long fingers. Since a seminal paper by Parlange and Hill [1976], gravity-driven unstable flow has often been a problem in which many soil physicists cut their teeth. During this time much progress has been made on when and why the flow becomes unstable, how it depends on the particular porous media and initial conditions, and how an understanding of unstable flow is intimately connected with new models of unsaturated flow. We start with the traditional fluid-fluid instability analysis originally formulated by Saffman and Taylor [1958].

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## Capillary effect on water table fluctuations in unconfined aquifers

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[1] Parlange and Brutsaert (1987) derived a modified Boussinesq equation to account for the capillary effect on water table dynamics in unconfined aquifers. Barry et al. (1996) solved this equation subject to a periodic boundary condition. Their solution shows significant influence of capillarity on water table fluctuations, which evolve to finite-amplitude standing waves at the high frequency limit. Here we propose a new governing equation for the water table, which considers both horizontal and vertical flows in an unsaturated zone of finite thickness. An approximate analytical solution for periodic water table fluctuations based on the new equation was derived. In agreement with previous results, the analytical solution shows that the unsaturated zone's storage capacity permits water table fluctuations to propagate more readily than predicted by the Boussinesq equation. Furthermore, the new solution reveals a capping effect of the unsaturated zone on both the amplitude and phase of the water table fluctuations as well as the water table overheight. Due to the finite thickness of the unsaturated zone, the capillary effect on water table fluctuations is modified mainly with reduced amplitude damping and phase shift.

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### 1. Introduction

[2] Oceanic oscillations produce water table fluctuations in coastal unconfined aquifers. As they propagate inland, the water table fluctuations are attenuated with increasing time lags. These fluctuations, representing basic characteristics of coastal groundwater, provide important information for understanding the properties and behavior of coastal aquifers, and have been subjected to numerous investigations [e.g., Parlange et al., 1984; Nielsen et al., 1997; Jiao and Tang, 1999; Li et al., 2000a; Li and Jiao, 2003; Jeng et al., 2005].

Additional supporting information may be found in the online version of this article.

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Although most previous research has been focused on tide-induced low-frequency water table dynamics [e.g., Nielsen, 1990; Li et al., 2000b, 2000c; Jeng et al., 2002], high-frequency water table fluctuations due to waves have also been studied [Li et al., 1997].

[3] Traditionally, models of water table fluctuations are based on the Boussinesq equation, which predicts increasing rates of amplitude damping with the frequency of the oceanic oscillations [e.g., Parlange et al., 1984; Nielsen, 1990]. According to these models, high-frequency waves would not induce water table fluctuations in coastal unconfined aquifers to any considerable distance inland, a result inconsistent with field observations. Li et al. [1997] found that consideration of capillarity explains the transmission of high-frequency water table fluctuations in coastal aquifers.

[4] Parlange and Brutsaert [1987] examined the capillary effect on water table dynamics. As the water table fluctuates, the pressure distribution above the water table varies, resulting in local water exchange across the water table. Parlange and Brutsaert [1987] modified the Boussinesq equation with an additional term to account for this mass transfer process. Barry et al. [1996] combined the approaches of Parlange et al. [1984] and Parlange and Brutsaert [1987]. They obtained and applied a depth-integrated model with capillarity incorporated to study the propagation of small-amplitude oscillations in an unconfined aquifer and derived an approximate analytical solution. Their results showed that the damping rate of the water table fluctuations reaches an asymptotic finite value as the forcing frequency on the boundary increases. In other words, damping effects on high-frequency water table fluctuations are bounded. Under the influence of capillarity,

## Acoustic waves in unsaturated soils

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[1] Seminal papers by Brutsaert (1964) and Brutsaert and Luthin (1964) provided the first rigorous theoretical framework for examining the poroelastic behavior of unsaturated soils, including an important application linking acoustic wave propagation to soil hydraulic properties. Theoretical developments during the 50 years that followed have led Lo et al., (2005) to a comprehensive model of these phenomena, but the relationship of its elasticity parameters to standard poroelasticity parameters measured in hydrogeology has not been established. In the present study, we develop this relationship for three key parameters, the Gassman modulus, Skempton coefficient, and Biot-Willis coefficient by generalizing them to an unsaturated porous medium. We demonstrate the remarkable result that well-known and widely applied relationships among these parameters for a porous medium saturated by a single fluid are also valid under very general conditions for unsaturated soils. We show further that measurement of the Biot-Willis coefficient along with three of the six elasticity coefficients in the model of Lo et al. (2005) is sufficient to characterize poroelastic behavior. The elasticity coefficients in the model of Lo et al. (2005) are sensitive to the dependence of capillary pressure on water saturation and its viscous-drag coefficients are functions of relative permeability, implying that hysteresis in the water retention curve and hydraulic conductivity function should affect acoustic wave behavior in unsaturated soils. To quantify these as-yet unknown effects, we performed numerical simulations for Dune sand at two representative wave excitation frequencies. Our results show that the acoustic wave investigated by Brutsaert and Luthin (1964) propagates at essentially the same speed during imbibition and drainage, but is attenuated more during drainage than imbibition. Overall, effects on acoustic wave behavior caused by hysteresis become more significant as the excitation frequency increases.

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### 1. Introduction

#### 1.1. Acoustic Waves and Poroelasticity

[2] The theory of poroelasticity, a term coined by *Geertsma* [1966], aims to provide a fundamental mathematical framework to describe the time-dependent coupling between applied stress and pore fluid pressure that occurs when acoustic waves pass through a porous medium. It is based on foundations laid down more than 70 years ago in a famous series of papers by *Biot* [1941, 1956, 1962], who was motivated initially by the classic geomechanical problem of soil consolidation. *Biot* [1956, 1962] presented a theory of dynamic poroelasticity that provides a general continuum-scale description of wave propagation through an elastic porous medium containing a single fluid, with

wave attenuation arising from fluid viscosity. In addition to the dilatational and shear waves like those observed in wave propagation through an isotropic elastic solid, *Biot* [1956, 1962] correctly predicted the existence of a second dilatational wave (“Biot slow wave”) caused by the coupled out-of-phase motions of the elastic porous framework and the viscous pore fluid.

[3] Extending the approach of *Biot* [1956], *Brutsaert* [1964] developed the first rigorous theoretical basis for modeling the effects of partial water saturation, such as occurs routinely in soils, on elastic wave behavior in unconsolidated porous media. As was the case in the pioneering study of *Brutsaert* [1964], most attempts to model acoustic wave propagation in porous media containing two immiscible fluids [*Garg and Nayfeh*, 1986; *Tuncay and Corapcioglu*, 1997; *Wei and Muraleetharan*, 2002] have neglected the possibility of coupling among the fluid phases and solid framework caused by differences between their accelerations (“inertial drag”), as opposed to coupling caused by differences between their velocities (“viscous drag”). Inertial drag was taken into account in the Biot-inspired model of elastic wave behavior in unsaturated porous media developed by *Berryman et al.* [1988], but their model was simplified significantly by imposing the condition that the capillary pressure between the two immiscible pore fluids remains constant during acoustic wave excitation.

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## Geomorphic signatures on Brutsaert base flow recession analysis

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[1] This paper addresses the signatures of catchment geomorphology on base flow recession curves. Its relevance relates to the implied predictability of base flow features, which are central to catchment-scale transport processes and to ecohydrological function. Moving from the classical recession curve analysis method, originally applied in the Finger Lakes Region of New York, a large set of recession curves has been analyzed from Swiss streamflow data. For these catchments, digital elevation models have been precisely analyzed and a method aimed at the geomorphic origins of recession curves has been applied to the Swiss data set. The method links river network morphology, epitomized by time-varying distribution of contributing channel sites, with the classic parameterization of recession events. This is done by assimilating two scaling exponents,  $\beta$  and  $b_G$ , with  $|dQ/dt| \propto Q^\beta$  where  $Q$  is at-a-station gauged flow rate and  $N(l) \propto G(l)^{b_G}$  where  $l$  is the downstream distance from the channel heads receding in time,  $N(l)$  is the number of draining channel reaches located at distance  $l$  from their heads, and  $G(l)$  is the total drainage network length at a distance greater or equal to  $l$ , the active drainage network. We find that the method provides good results in catchments where drainage density can be regarded as spatially constant. A correction to the method is proposed which accounts for arbitrary local drainage densities affecting the local drainage inflow per unit channel length. Such corrections properly vanish when the drainage density become spatially constant. Overall, definite geomorphic signatures are recognizable for recession curves, with notable theoretical and practical implications.

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### 1. Introduction

[2] Groundwater is the main contributor of a river catchment's base flow whose predictability during recession events is of crucial importance for water resource management. Recession curves have been widely studied in the past and their characteristics used to establish basin-scale parameters (see *Tallaksen* [1995] for a review). In particular, *Brutsaert and Nieber* [1977] analyzed daily discharge values of six basins in the Finger Lakes region of the north-eastern US and proposed an analytical tool to characterize the recession flow based on the description of the discharge

change rate  $dQ/dt$  as a function of the discharge  $Q$ . Unlike many nonlinear recession flow models, this method avoids the knowledge of the precise beginning of the recession event which can be difficult to evaluate due to the continuous nature of streamflow measurements. The main feature of their method is the comparison of the observations with analytical solutions of the Boussinesq equation for an unconfined rectangular aquifer under particular boundary conditions. Two exact solutions of the Boussinesq equation [*Boussinesq*, 1904; *Polubarinova-Kochina*, 1962] and an approximated linearized solution [*Boussinesq*, 1903] can be expressed in the form:

$$\frac{dQ}{dt} = -kQ^\beta \quad (1)$$

where  $\beta$  and  $k$  are constants depending on the flow regime considered. In order to avoid contributions from relatively fast subsurface flow, overland flow, and evapotranspiration, *Brutsaert and Nieber* [1977] recommended the use of the lower envelope of the point cloud in the  $\ln(-dQ/dt)$  versus  $\ln Q$  plot, corresponding to the slowest recession rate. Based on their study, they identified two typical values of  $\beta$ , describing the rate of decline in streamflow recessions:  $\sim 1.5$  for low  $Q$  (long-term response) and  $\sim 3$  for high  $Q$  (short-term response). Moreover, some parameters of the

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## Polynomial approximate solutions of a generalized Boussinesq equation

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[1] A generalized Boussinesq equation, the porous medium equation, was analyzed for a semi-infinite initially dry unconfined aquifer. The boundary conditions were a power-law head condition at the inlet boundary and a zero-head condition at infinity. Quadratic and cubic polynomial approximate solutions to the equation were derived. These approximate solutions replicate known exact solutions of the Boussinesq and porous medium equations. The approximate solutions were also compared to numerical solutions of the generalized Boussinesq equation computed using a method of Shampine. It was found that the solutions are easy to use and they have sufficient accuracy to be useful in practical applications, such as when the hydraulic conductivity is a power-law function of elevation.

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### 1. Introduction

[2] The Boussinesq equation is used to model groundwater flow in unconfined aquifers when the Dupuit-Forchheimer assumption is valid. The porous medium equation is a generalization of the Boussinesq equation that can be derived when the hydraulic conductivity is a power-law function of elevation [Rupp and Selker, 2005]. For example, such situations appear during the filtration of water in concretes [Lockington et al., 1999] and in flows in forest soils [Beven, 1982]. The porous medium equation also models the flow of polytropic gases, gases where the pressure in the gas is a power-law function of density, in porous media [Barenblatt, 1952]. Moreover, the porous medium equation is a special case of a more general nonlinear equation appearing in mathematical hydraulics [Daly and Porporato, 2004].

[3] Lockington et al. [2000] analyzed the problem of flow in a one-dimensional unconfined semi-infinite horizontal aquifer. The aquifer was assumed to be initially dry with head as a power-law function of time at the inlet. The head at infinity was zero. The flow was modeled using the Boussinesq equation. Dimensional analysis was used to convert the problem to a boundary value problem for an ordinary differential equation. A quadratic approximate solution to this problem was derived. The approximate solution was in agreement with the numerical solution obtained using the method of Shampine [1973]. Furthermore, the quadratic approximate solution reproduced

two known exact solutions of the Boussinesq equation [Barenblatt, 1952].

[4] Telyakovskiy et al. [2002] extended the work of Lockington et al. [2000] by constructing a cubic approximate solution to the same boundary value problem. This solution improved the quadratic solution of Lockington et al. [2000] and reproduced the two known exact solutions. The method was later applied to other types of boundary conditions in Telyakovskiy and Allen [2006] and Telyakovskiy and Kurita [2007].

[5] In this note, we apply the methods of Lockington et al. [2000] and Telyakovskiy et al. [2002] to model one-dimensional, unconfined groundwater flow in a horizontal aquifer using the porous medium equation

$$\partial_t h = a \partial_x^2 h^k, \quad (x, t) \in (0, \infty) \times (0, \infty) \quad \text{and} \quad k > 1. \quad (1)$$

[6] Such equations occur when hydraulic conductivity is a power-law function of water head,  $K(h) = K_0 h^m$ . In equation (1),  $a$  depends on properties of the porous medium and fluid,  $h$  is the head,  $k$  is related to the exponent in the conductivity power law by  $k = m + 1$ ,  $x$  is the horizontal distance from the left boundary and  $t$  is time.

[7] We assume that the aquifer is initially dry,

$$h(x, 0) = 0 \quad \text{for} \quad x > 0, \quad (2)$$

and that a power-law describes the head at the inlet boundary,

$$h(0, t) = \sigma t^\alpha, \quad \sigma > 0 \quad \text{and} \quad -\frac{1}{k+1} \leq \alpha < \infty. \quad (3)$$

[8] The parameters  $\alpha$  and  $\sigma$  are constants. The value of  $\alpha$  cannot be less than  $-(k+1)^{-1}$ , otherwise the water level at the inlet will be decreasing faster than the water level in the aquifer. In this case, the water would be discharged from the aquifer into the body of water, which is not the recharge situation considered here. For example, for flow

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## Evaluating the influence of watershed moisture storage on variations in base flow recession rates during prolonged rain-free periods in medium-sized catchments in New York and Illinois, USA

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[1] When  $dQ/dt-Q$  plots of stream recession are constructed for individual recession events, the slopes of the  $dQ/dt-Q$  curves are near constant in log space, but the intercepts vary seasonally. Because the intercepts increase during the summer (indicating an increase in the recession rate at a given discharge), it has often been assumed that increased evapotranspiration (ET) leads to increased recession rates. To test this assumption, we related the intercepts of  $dQ/dt-Q$  curves from 72 recession events to the concurrent ET and watershed moisture storage as determined from the National Center for Environmental Prediction (NCEP) North American Regional Reanalysis (NARR) data set. The analysis suggests that at least for the nine watersheds from Illinois and New York we studied, shifts in recession rate during prolonged rain-free periods had little linkage to concurrent ET. Instead, we observe that the shifting has a moderately strong linkage to watershed moisture storage during the recession event. While this seeming lack of dependence on ET during these prolonged rain-free periods is not necessarily reflective of more normal conditions, we suggest it provides some insight into underlying subsurface controls at the watershed scale. In particular, we hypothesize that the shift in intercept in  $dQ/dt-Q$  curves results from spatial heterogeneities in watershed surficial geology; under dryer conditions near-stream subsurface zones with high hydraulic conductivities contribute most streamflow but under wetter conditions subsurface zones in upland areas with lower hydraulic conductivities also contribute.

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### 1. Introduction

[2] Plots relating the rate of change in stream discharge ( $dQ/dt$ ) versus the concurrent mean discharge ( $Q$ ) have long been used to make inferences into the aquifer properties of watersheds [Brutsaert and Nieber, 1977]. New work has investigated the possibility of considering  $dQ/dt-Q$  plots more broadly as a means to infer the storage-discharge relationship [Kirchner, 2009] or changes in the contributing channel network of a watershed [Biswal and Marani, 2010]. Traditionally, a key limitation to interpreting  $dQ/dt-Q$  plots has been the large amount of scatter

among the data points. To date, different investigators have considered different approaches for dealing with the scatter: looking at the envelope of the data cloud, fitting a best-fit line through the middle of the data cloud, or binning data with similar  $Q$  and fitting a line to the collection of mean  $dQ/dt$  values within each bin. As an alternative to these approaches, Shaw and Riha [2012] and Biswal and Marani [2010] examined individual recession events within the data cloud. When viewed in this way, it becomes apparent that the data cloud largely consists of multiple individual events that shift seasonally (Figure 1). Relative to the extrapolated curve of  $\log(-dQ/dt)$  versus  $\log(Q)$  during wet months (the bold black line in Figure 1), the  $\log(-dQ/dt)$  versus  $\log(Q)$  curves in the summer months shift upward. Thus, there appears to be more structure to the data cloud than previously thought, if the shifts can be explained.

[3] Shaw and Riha [2012] hypothesized that the shifts were due to variations in evapotranspiration (ET) during different recession events. Physically, increasing ET could be assumed to concurrently draw down the water available for recession flow and thus make the recession rate faster, as evident by a larger intercept value of event  $dQ/dt-Q$  curves during the summer time. This conclusion is consistent with much of the dominant thinking regarding controls

Additional supporting information may be found in the online version of this article.

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## Pore scale consideration in unstable gravity driven finger flow

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[2] To explain the dynamic behavior of the matric potential at the wetting front of gravity driven fingers, we take into account the pressure across the interface that is not continuous and depends on the radius of the meniscus, which is a function of pore size and the dynamic contact angle  $\theta_d$ .  $\theta_d$  depends on a number of factors including velocity of the water and can be found by the Hoffman-Jiang equation that was modified for gravity effects. By assuming that water at the wetting front imbibes one pore at a time, realistic velocities are obtained that can explain the capillary pressures observed in unstable flow experiments in wettable and water repellent sands.

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### 1. Introduction

[3] Fingering flow initiated by gravity has been studied for over 40 years. It is a special case of water infiltration in a porous medium where equilibrium capillary pressure-saturation conditions are not maintained at the wetting front [Hsu and Hilpert, 2011; Mumford and O'Carroll, 2012; DiCarlo 2013]. At the wetting front where fluid saturation changes rapidly, measurements of finger flow behavior by among others Liu *et al.* [1993] and DiCarlo [2004, 2007] show that capillary pressure is a function of the rate of change of saturation, referred to as dynamic capillary pressure [Hsu and Hilpert, 2011]. This discrepancy in predicting wetting front behavior between the static and dynamic approach is especially great when the static contact angle (also called equilibrium contact angle) is intermediate between  $0^\circ$  and  $90^\circ$  [O'Carroll *et al.*, 2010]. The latter might explain why unstable fingering flow is observed in experiments in silica sand with a contact angle between  $30^\circ$  and  $60^\circ$  [Extrand and Kumagai, 1997; Romano, 2006; Schroth *et al.*, 1996] or extremely dry soil [Nektarios *et al.*, 1999].

[4] An agreed upon explanation of the dynamic behavior of the matric potential at the wetting front remains elusive despite investigations by multiple laboratories [O'Carroll *et al.*, 2010]. Most approaches to model the dynamic matric

potential assume that the pressure field at the wetting front is continuous (see DiCarlo 2013, this special issue, for an overview of past approaches). The continuum approach of Hsu and Hilpert [2011] and Hilpert [2012] in which the change in dynamic contact angle and resulting matric potential at the fluid-air interface is related to the velocity of the front seems to be particularly promising. Other modelers employed pore network models that assume that the pressure field can be discontinuous, such that water does not flow through all the pores at the wetting front at the same time [Joekar-Niasar and Hassanizadeh, 2012; DiCarlo, 2013, this special issue]. Very interesting and relevant are findings of Moebius and Or [2012] and Moebius *et al.* [2012] who recorded water moving through pores in packed glass beads with high speed camera and acoustics. They found that all menisci increase in size at the front until water broke through in one pore followed by a decrease in size for all other menisci. Finally, Bayer *et al.* [2013, this special issue] also assumed a discontinuous pressure field at the wetting front.

[5] In the Bayer *et al.* [2013] paper, we argue that the key in explaining finger formation under gravity is that the pressure across the wetting front is discontinuous and determined by the shape of the dynamic (or nonequilibrium) contact angle between the meniscus and the sand grain. Furthermore, we showed that the Hoffman equation relating dynamic contact angle to velocity of moving contact line could be applied in porous media in which gravity was the driving force. By employing a contact line velocity in Hoffman's relationship, we could estimate the matric potential at the wetting front of water moving down in small columns similar to those used in the Geiger and Durnford [2000] experiments with the same precision as Hsu and Hilpert [2011], but with one fitting parameter less because we employed the Hoffman equation. The velocity was calculated by assuming that the imposed flux passed through one or several pores each instant while the other pores had no flow. The two fitting parameters in Bayer *et al.* [2013] were (1) a grain size dependent static contact angle, and (2) the number of pores imbibing the imposed flux each

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## Modeling effect of initial soil moisture on sorptivity and infiltration

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[1] A soil's capillarity, associated with the parameter sorptivity, is a dominant control on infiltration, particularly at the onset of rainfall or irrigation. Many mathematical models used to estimate sorptivity are only valid for dry soils. This paper examines how sorptivity and its capillary component (as wetting front potential) change with initial degree of saturation. We capture these effects with a simple modification to the classic Green-Ampt model of sorptivity. The modified model has practical applications, including (1) accurately describing the relative sorptivity of a soil at various water contents and (2) allowing for quantification of a soil's saturated hydraulic conductivity from sorptivity measurements, given estimates of the soil's characteristic curve and initial water content. The latter application is particularly useful in soils of low permeability, where the time required to estimate hydraulic conductivity through steady-state methods can be impractical.

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### 1. Introduction

[2] Because infiltration affects water availability for vegetation, groundwater recharge, overland flow, and solute transport, it has been the focus of considerable study over the previous century [e.g., *Green and Ampt*, 1911; *Philip*, 1957b; *Wooding*, 1968; *Brutsaert*, 1977]. Under normal conditions, gravity and capillarity drive vertical infiltration, whereas capillarity alone drives horizontal infiltration [*Philip*, 1957b].

[3] Under constant head conditions, one- and three-dimensional vertical infiltration into a uniform soil has been adequately described using *Philip's* [1957b] two-term approximation:

$$I = S\sqrt{t} + Ct \quad (1)$$

where  $I$  is cumulative infiltration over time  $t$  and  $S$  is the soil sorptivity. For one-dimensional vertical infiltration,  $C$  is proportional to the soil's saturated hydraulic conductivity ( $K_s$ ). The ratio  $C/K_s$  is  $\leq 1$ , depending on soil type and soil moisture [*Philip*, 1990], with proposed ranges of  $1/3 \leq C/K_s \leq 2/3$  [*Fuentes et al.*, 1992] or  $0.3 \leq C/K_s \leq 0.4$  [*Philip*, 1990]. In the case of three-dimensional infiltration,  $C$  incorporates both saturated hydraulic conductivity and sorptivity [*Smettem et al.*, 1995; *Touma et al.*, 2007].

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[4] At early times (i.e.,  $t \ll S^2/C^2$ ) sorptivity dominates the infiltration behavior, and for very early times ( $t \rightarrow 0$ ) the second term on the right hand side may be neglected [*White et al.*, 1992]. Conversely, the second term dominates as time increases, subject to the limit of  $t = S^2/C^2$ , when the series expansion from which equation (1) was derived is no longer accurate. Alternate expressions have been developed to describe long-time (steady-state) infiltration behavior [*Philip*, 1957a, 1957b; *Wooding*, 1968; *Haverkamp et al.*, 1994], which lend themselves to estimations of  $K_s$ . However, the time required to reach late-time or quasi-steady state conditions may be impractical, particularly for soils with low hydraulic conductivity, and assumptions of homogeneity are typically violated for long infiltration experiments.

[5] Infiltration typically occurs over intermediate or transient timescales (neither exclusively early- nor late-time) and is three-dimensional. One such example is infiltration from an axisymmetric single ring source, which can provide a rapid and low-cost measurement of soil hydraulic properties [*Braud et al.*, 2005]. However, interpretation of these infiltration tests often requires that the  $S$  and  $C$  terms both be considered. Methods to differentiate between sorptivity and saturated hydraulic conductivity for such infiltration conditions have been proposed [*Smiles and Knight*, 1976; *Smettem et al.*, 1995; *Vandervaere et al.*, 2000], but may be inadequate for estimating small  $K_s$  values [*Smettem et al.*, 1995].

[6] Sorptivity represents the soil's ability to draw water [*Philip*, 1957b; *Touma et al.*, 2007], which is a function of the capillarity (the driving force) and the soil's hydraulic conductivity (the dissipation). This dual-dependence is evident in *Parlange* [1975]'s precise solution for sorptivity (as modified for positive ponded conditions by *Haverkamp et al.* [1990]):

$$S^2 = 2K_s(\theta_s - \theta_r)(1 - \Theta_0)h_{surf} + (\theta_s - \theta_r) \int_{h_0}^0 (1 + \Theta - 2\Theta_0)K(h)dh \quad (2)$$

## The importance of hydraulic groundwater theory in catchment hydrology: The legacy of Wilfried Brutsaert and Jean-Yves Parlange

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[1] Based on a literature overview, this paper summarizes the impact and legacy of the contributions of Wilfried Brutsaert and Jean-Yves Parlange (Cornell University) with respect to the current state-of-the-art understanding in hydraulic groundwater theory. Forming the basis of many applications in catchment hydrology, ranging from drought flow analysis to surface water-groundwater interactions, hydraulic groundwater theory simplifies the description of water flow in unconfined riparian and perched aquifers through assumptions attributed to Dupuit and Forchheimer. Boussinesq (1877) derived a general equation to study flow dynamics of unconfined aquifers in uniformly sloping hillslopes, resulting in a remarkably accurate and applicable family of results, though often challenging to solve due to its nonlinear form. Under certain conditions, the Boussinesq equation can be solved analytically allowing compact representation of soil and geomorphological controls on unconfined aquifer storage and release dynamics. The Boussinesq equation has been extended to account for flow divergence/convergence as well as for nonuniform bedrock slope (concave/convex). The extended Boussinesq equation has been favorably compared to numerical solutions of the three-dimensional Richards equation, confirming its validity under certain geometric conditions. Analytical solutions of the linearized original and extended Boussinesq equations led to the formulation of similarity indices for baseflow recession analysis, including scaling rules, to predict the moments of baseflow response. Validation of theoretical recession parameters on real-world streamflow data is complicated due to limited measurement accuracy, changing boundary conditions, and the strong coupling between the saturated aquifer with the overlying unsaturated zone. However, recent advances are shown to have mitigated several of these issues. The extended Boussinesq equation has been successfully applied to represent baseflow dynamics in catchment-scale hydrological models, and it is currently considered to represent lateral redistribution of groundwater in land surface schemes applied in global circulation models. From the review, it is clear that Wilfried Brutsaert and Jean-Yves Parlange stimulated a body of research that has led to several fundamental discoveries and practical applications with important contributions in hydrological modeling.

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## An extension of Miller scaling to scale sorptivity by contact angle

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[1] This study sheds light on the limitations of using  $[(\cos \theta)^{1/2}]$  to scale sorptivity by contact angle while reaffirming its scaling by geometrical Miller scaling factor  $(\lambda^{1/2})$ . The sorptivity for uniform and nonuniform (wavy) capillary tubes was determined by a mathematical model that includes the effect of inertia and dynamic contact angle. Given that real porous media are preferably represented by a bundle of nonuniform rather than uniform capillary tubes, the relationship between sorptivity and contact angle was examined for different combinations of contact angles and capillary tube degrees of waviness. A general relationship of  $S = f[\cos \theta]^\beta$  (with  $\beta \leq 1/2$ ) was found. The deviation of  $\beta$  from  $1/2$  (associated with uniform capillary tubes) increased with contact angle and capillary waviness increase. Zero sorptivity was obtained even for wettable capillaries,  $\theta < 90^\circ$ , a phenomenon that has been generally associated with hydrophobic capillaries ( $\theta \geq 90^\circ$ ). Contact angle and nonuniform pore structure had a synergistic effect on sorptivity. Capillary nonuniformity per se diminished sorptivity but its synergy with contact angle markedly magnified this reduction. Thus, following the sorptivity impact on finger width, it is rational to assume that larger-than-zero contact angles are involved in the formation of narrow fingers with an abrupt change between the inner wet and surrounding dry areas.

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### 1. Introduction

[2] The size of fingers formed by unstable flow, their width in particular, has been a topic of interest for many years. Models used to determine these variables have mostly emanated from the theory of unstable flow in porous media. *DiCarlo* [2013] describes the evolution of the theory used to study unstable and fingered flow which initiated with the linear stability analysis of *Saffman and Taylor* [1958] and *Chouke et al.* [1959]. The theory categorizes stability/instability by the interplay of gravitational (destabilizing) and viscosity and capillary (stabilizing) forces. Flow is considered stable when a small perturbation introduced into the advancing sharp front that separates two immiscible fluids decays, and unstable when this small perturbation grows, forming a finger that dominates the displacement process. While *Saffman and Taylor* [1958] considered viscosity and gravity, *Chouke et al.* [1959] included capillary forces in the stabilizing effect in the form of an effective surface tension ( $\sigma_*$ ) at the macroscopic interface on Darcy's scale. The finger width (2D) or diameter (3D) is predicted by half of the critical wavelength at which the

capillary forces act to stabilize the perturbations to the wetting front. Therefore, the minimum finger size,  $d_c$ , for a particular porous medium and applied flux is predicted by *Chouke et al.* [1959] as

$$d_c = a\sqrt{\sigma_*k} \left( \frac{1}{U\theta_s(\mu_2 - \mu_1) + kg(\rho_1 - \rho_2)} \right)^{1/2} \quad (1)$$

[3] *Parlange and Hill* [1976] considered the capillary forces as parameterized by soil water sorptivity ( $S$ ) in the linear stability analysis used to calculate the minimum finger size

$$d_c = a \frac{S^2(\mu_1 + \mu_2)}{2(\theta_s - \theta_i)} \left( \frac{1}{U\theta_s(\mu_1 - \mu_2) + kg(\rho_1 - \rho_2)} \right) \quad (2)$$

where  $\mu$  and  $\rho$  are the fluids' dynamic viscosity and density, respectively. The sorptivity is evaluated between  $\theta_s$ , the saturated moisture content, and  $\theta_i$ , the initial moisture content, and  $a$ , a constant equal to  $\pi$  for a 2D system and 4.8 for a 3D system [*Glass et al.*, 1991]. Application of equation (2) to the air/water system where the density and viscosity of air are negligible with respect to those of water, and defining a dimensionless system flux ratio ( $R_s = q_s/K_s$ ) led to the following simplifications [*Glass and Nicholl*, 1996]

$$d_c = \alpha \frac{S^2}{2K_s(\theta_s - \theta_i)} \left( \frac{1}{1 - R_s} \right) \quad (3)$$

[4] The stability criterion of *Saffman and Taylor* [1958], weighing viscous against gravitational forces, is contained

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# A fully stochastic approach bridging the microscopic behavior of individual microorganisms with macroscopic ensemble dynamics in surface flow networks

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[1] Prediction of microbial surface water contamination is a formidable task because of the inherent randomness of environmental processes driving microbial fate and transport. In this article, we develop a theoretical framework of a fully stochastic model of microbial transport in watersheds, and apply the theory to a simple flow network to demonstrate its use. The framework bridges the gap between microscopic behavior of individual microorganisms and macroscopic ensemble dynamics. This scaling is accomplished within a single mathematical framework, where each microorganism behaves according to a continuous-time discrete-space Markov process, and the Markov behavior of individual microbes gives rise to a nonhomogeneous Poisson random field that describes microbial population dynamics. Mean value functions are derived, and the spatial and temporal distribution of water contamination risk is computed in a straightforward manner.

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## 1. Introduction

[2] Fecal contamination is the leading cause of surface water impairment in the US [USEPA, 2008], and fecal pathogens are capable of triggering massive outbreaks of gastrointestinal disease. The most notable outbreak took place in Milwaukee, WI in 1993; it affected over 400,000 people, contributed to more than 100 deaths and was *not predicted* by water resources management [MacKenzie *et al.*, 1994]. It is well known that the difficulty in prediction of water contamination has its roots in the stochastic variability of fecal pathogens in the environment, and in the complexity of environmental systems [e.g., Bertuzzo *et al.*, 2007; Tyrrel and Quinton, 2003]. The environmental parameters that have a high degree of influence on surface water flow and contamination include rainfall, hydraulic conductivity, soil composition, surface roughness, slope, and vegetation among others. The variability of some of them seems to evolve randomly and cannot be described deterministically at the current level of knowledge, even if the underlying phenomena were thoroughly understood [Kottegoda, 1980]. Instead, posing them in a probabilistic

framework allows for a quantitative, rational treatment [Clarke, 1998; Freeze, 1980].

[3] The complexity of environmental systems and our lack of knowledge about the physical, biological, and chemical aspects of microbial fate and transport is yet another significant barrier in accurate prediction of water contamination events. For instance, general trends in microbial transport can be captured using standard hydrodynamic models that do not incorporate specifics of microbial behavior. This approach, however, leaves dynamic processes during and poststorm events unexplained [Cho *et al.*, 2010; Hellweger and Masopust, 2008]. This is problematic because many water contamination events and associated waterborne disease outbreaks take place after storms [Curriero *et al.*, 2001]. For example, a specific aspect of storm and poststorm events that has recently been addressed by several groups is microbial-soil sediment interactions in general, and deposition and resuspension of microbes from bottom sediments in particular. It appears that streambeds serve as a significant, albeit previously unaccounted, source of microbes in surface waters, particularly during and after storm events, and inclusion of these sources, along with resuspension and deposition of microbes during transport, improves model predictions [see, e.g., Cho *et al.*, 2010; Jamieson *et al.*, 2005; Kim *et al.*, 2010; Pandey *et al.*, 2012; Yeghiazarian *et al.*, 2006, among others]. This view necessitates the inclusion of sediment dynamics in microbial transport models. The overall conclusion, therefore, is that in order to accurately predict microbial surface water contamination events, physics-based stochastic models are needed.

[4] Most models of microbial transport in surface water, however, including those developed by regulatory

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