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Improving the representation of hydrologic processes in Earth System Models

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Key Points:
- Land model development can benefit from recent advances in hydrology
- Accelerating modeling advances requires comprehensive benchmarking activities
- Stronger collaboration is needed between the hydrology and ESM modeling communities

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Abstract Many of the scientific and societal challenges in understanding and preparing for global environmental change rest upon our ability to understand and predict the water cycle change at large river basin, continent, and global scales. However, current large-scale land models (as a component of Earth System Models, or ESMs) do not yet reflect the best hydrologic process understanding or utilize the large amount of hydrologic observations for model testing. This paper discusses the opportunities and key challenges to improve hydrologic process representations and benchmarking in ESM land models, suggesting that (1) land model development can benefit from recent advances in hydrology, both through incorporating key processes (e.g., groundwater-surface water interactions) and new approaches to describe multiscale spatial variability and hydrologic connectivity; (2) accelerating model advances requires comprehensive hydrologic benchmarking in order to systematically evaluate competing alternatives, understand model weaknesses, and prioritize model development needs, and (3) stronger collaboration is needed between the hydrology and ESM modeling communities, both through greater engagement of hydrologists in ESM land model development, and through rigorous evaluation of ESM hydrology performance in research watersheds or Critical Zone Observatories. Such coordinated efforts in advancing hydrology in ESMs have the potential to substantially impact energy, carbon, and nutrient cycle prediction capabilities through the fundamental role hydrologic processes play in regulating these cycles.

1. Introduction

The stores and fluxes of water near the land surface modulate nearly all processes at the land-atmosphere interface. For example, evapotranspiration affects weather and climate through its influence on boundary layer dynamics and thermal dynamics. The stores and fluxes of water on land also influence ecosystem structure and functioning, carbon, nutrient, and other biogeochemical cycles, and dictate freshwater availability for societies. Many of the scientific and societal challenges to understand and prepare for global environmental change rest upon our ability to understand and predict water cycle changes at large river basin, continent, and global scales [*Eagleson*, 1986].

Simulations of global environmental change are now performed using Earth System Models (ESMs), which are designed to capture the key interactions among the components of the Earth System including the atmosphere, ocean, land, and sea ice [*Flato*, 2011; *Dunne et al.*, 2012; *Hurrell et al.*, 2013]. ESMs build on global climate model platforms but substantially extend their capabilities by representing a diverse set of physical, chemical, and biological interactions across multiple space and time scales [*Hurrell et al.*, 2013]. Here we focus on the hydrologic processes in the so-called "land models"—the land component of ESMs. Because terrestrial water stores and fluxes are strong modulators of energy and biogeochemical stores on land and their flux exchange with the atmosphere and the oceans, adequately capturing hydrologic processes in the ESM land models has important implications to our ability to predict global environmental change.

The last 50 years have seen remarkable advances in the capabilities of land models to simulate land surface influences on weather and climate. Since the very first attempt to incorporate land surface processes in climate models [*Manabe et al*., 1965; *Manabe*, 1969], land models have evolved to include sophisticated representation of biogeophysical, biogeochemical, and hydrologic processes [*Sellers et al*., 1997; *Pitman*, 2003; *van den Hurk et al*., 2011]. Many hydrologists have been actively involved in, and contributed toward, water cycle advancement in ESMs, and several such land models have been applied to studying the water cycle change at large river basin, continent and global scales. This two-way flow of knowledge and tools have benefited both the ESM community and those hydrologists engaged in large-scale water cycle research. As *Water Resources Research* celebrates its Fiftieth anniversary, it is important to reflect on the past advances in water cycle representation in ESM land models, and consider if the collective wisdom and knowledge from the hydrologic sciences could be more fully utilized to inform and advance the representation of the terrestrial water cycle in ESMs.

In this paper, we juxtapose modeling advances in the land modeling community with advances in the hydrologic sciences to address the following two questions:

1. What are the hydrologic stores and fluxes that are known to be essential to predicting large-scale water, energy, and biogeochemical fluxes but not yet adequately accounted for in ESMs? Put differently, what are the key opportunities to improve the representation of hydrologic processes in land models?
2. How can we evaluate model improvements? What are the practical steps that our community can take so that we can meaningfully and transparently test, document, and benchmark model performance, and set milestones moving forward as a community?

We do not intend to fully answer these questions here, but to provide our perspective of the past progress and lessons learned, building on the extensive literature of Model Intercomparison Experiments and benchmarking activities (see *van den Hurk et al*. [2011] for a recent review), along with a recent synthesis of modeling approaches in the hydrologic sciences [*Clark et al*., 2015a, 2015b]. We also do not intend to provide a comprehensive review of the representation of all terrestrial hydrologic processes in all models designed for many different applications. Our emphasis here is the hydrologic processes that are well recognized to be the most basic (e.g., lateral convergence, two-way groundwater-surface water exchange) and most relevant to the large-scale Earth System functioning (e.g., regulating the global carbon cycle). We also emphasize the critical need to represent hydrologic processes explicitly and mechanistically (not empirically) so that we can predict future water cycle changes forced by, and forcing, global energy and biogeochemical cycle changes. Lastly, we emphasize the need for parsimonious (parameter-wise) and efficient (computation-wise) ways to represent these essential hydrologic processes in the ESM context.

To keep the paper tractable, we focus primarily on below ground hydrologic processes and explicitly consider the representation of the storage and transmission of water in soils, infiltration, root water uptake, groundwater dynamics, stream-aquifer interactions, and channel/floodplain routing. These stores and fluxes are arguably among the weakest parts of current land models. We also review model representations of spatial heterogeneity and briefly discuss the impact of humans on the terrestrial water cycle. Important hydrologic processes, such as canopy and snow hydrology, the role of land cover and land use change on the water cycle, and interactions between hydrology and biogeochemistry, are not reviewed here.

The remainder of this paper is organized as follows. In section 2, we provide the scientific motivation for improving the representation of hydrologic processes in land models. In section 3, we contrast the current representation of hydrologic processes in modern land models (Table 1) with the modeling approaches used in the hydrologic sciences. Our goal for this (rather focused) review is to identify key opportunities to improve the representation of hydrologic processes in the land models of ESMs. In section 4, we discuss the work needed to accelerate modeling advances; in particular, we discuss the need for comprehensive hydrologic benchmarking activities that systematically evaluate competing modeling alternatives, understand model weaknesses, and prioritize model development needs. Finally, in section 5, we present concluding remarks.

Table 1. Land Models Reviewed in This Study

Model	Name	References
CABLE	The Community Atmosphere Biosphere Land Exchange model	*Kowalczyk et al.* [2006] and *Wang et al.* [2011]
Catchment	The Catchment model	*Ducharne et al.* [2000] and *Koster et al.* [2000]
CLM	The Community Land Model	*Niu et al.* [2007], *Zeng and Decker* [2009], *Oleson et al.* [2010], and *Lawrence et al.* [2011]
TESSEL	The Tiled ECMWF Scheme for Surface Exchanges over Land	*Balsamo et al.* [2009], *Balsamo et al.* [2011], *Pappenberger et al.* [2012], and *ECMWF* [2014]
JULES	The Joint UK Land Environment Simulator	*Cox et al.* [1999], *Gedney and Cox* [2003], *Clark and Gedney* [2008], and *Best et al.* [2011]
LEAF	The Land Ecosystem-Atmosphere Feedback model	*Walko et al.* [2000], *Fan et al.* [2007], Miguez-*Macho et al.* [2007], and *Miguez-Macho and Fan* [2012a,b]
LM3	Land Model 3	*Milly et al.* [2014] and *Subin et al.* [2014]
MATSIRO	The Minimal Advanced Treatments of Surface Interaction and Runoff	*Takata et al.* [2003], *Yeh and Eltahir* [2005a,b], and *Koirala et al.* [2014]
Noah	The Noah land surface model	*Chen et al.* [1996], *Schaake et al.* [1996], *Koren et al.* [1999], *Chen and Dudhia* [2001], and *Ek et al.* [2003]
ORCHIDEE	The Organizing Carbon and Hydrology in Dynamic Ecosystems model	*de Rosnay et al.* [2002], *Verant et al.* [2004], *Ngo-Duc et al.* [2007], *d'Orgeval et al.* [2008], and *Campoy et al.* [2013]
VIC	The Variable Infiltration Capacity model	*Liang et al.* [1994], *Liang et al.* [1996], *Lohmann et al.* [1998b], and *Nijssen et al.* [2001]

2. Scientific Motivation

The current rudimentary representation of hydrologic processes in ESM land models limits our capability to simulate land-atmosphere fluxes and biogeochemical cycles, which, in turn, limits our ability to predict land-atmosphere processes and feedbacks across multiple time (minutes to centuries) and space (points to continents) scales. Incomplete treatment of terrestrial hydrological processes also impedes our ability to properly constrain model performance using new observational capabilities such as groundwater and remotely sensed data. A concerted effort to improve the representation of hydrologic processes in land-surface models will accelerate science advances in a number of key areas. Some of the science questions that motivate the overarching ideas presented in this review include:

1. How does the space-time variability in hydrologic stores (canopy, snowpack, soil moisture, groundwater, rivers and floodplains, lakes and reservoirs, and wetlands) affect land-atmosphere fluxes on time scales from minutes to seasons? How does the hydrologic modulation of land-atmosphere fluxes affect weather and climate, including fronts, storm tracks, monsoons, and extremes? To what extent do advances in model representation of the land branch of the water cycle improve regional predictions of weather and climate across multiple time and space scales?
2. How will the hydrologic stores and fluxes respond to global environmental change under natural (e.g., climate variability) and anthropogenic (e.g., land use change, irrigation) forcings? How will the water cycle response affect biological productivity and functioning in terrestrial and aquatic ecosystems? How will the hydrologic and ecosystem responses feedback on the climate system, and through what physical-biogeochemical pathways? What are the globally salient mechanisms of these responses and feedbacks?
3. Given that groundwater and lateral flow are largely missing in ESM land models, what is the role of the groundwater, with its large storage and slow response to climate forcing, in buffering water stress in terrestrial and aquatic ecosystems and human societies? How will groundwater resources respond to climate change and mounting human pressure? How will the globally diminishing groundwater observed by GRACE affect dry-season streamflow, groundwater-supported ecosystem productivity, and water quality (e.g., stream temperature)?
4. What parts of the world will likely experience water stress, posing threats to ecosystem sustainability and water and food security for human populations? Can we quantify water resource vulnerability in the context of climate change and human development uncertainty?
5. How will warming affect the Pan-Arctic permafrost and peatlands? How will the projected large-scale thawing and draining affect Arctic hydrological conditions, thereby impacting how and when thawed organic matter is decomposed and released to the atmosphere as carbon dioxide or methane?

These critical science questions cannot be comprehensively addressed with the current generation of ESM land models. Addressing them requires additional targeted development, capitalizing on the progress in modeling capabilities in the hydrologic sciences. The next section outlines some key opportunities to advance representation of hydrologic processes in land models necessary to begin to address these critical science questions.

3. Opportunities to Improve the Representation of Hydrologic Processes in Land Models

The opportunities to improve the representation of hydrologic processes in land models fall into three main categories. First, there are opportunities to directly improve the model representation of individual hydrologic processes. Work in this area can improve simulations of soil moisture dynamics, groundwater dynamics, and, ultimately, the space-time variability of plant available water and its impact on biogeochemical cycles. Second, there are opportunities to improve representations of how interlinked hydrologic-biophysical-biogeochemical processes affect land-atmosphere fluxes at larger spatial scales. The key aspects of this second element are the approaches used to characterize spatial heterogeneity and hydrologic connectivity. Third, there are opportunities to substantially improve model representations of human impacts on the terrestrial hydrologic cycle. The following sections review model development opportunities in each of these three areas.

3.1. Representation of Hydrologic Processes

Table 2 summarizes the key differences in the representation of hydrologic processes among 11 current generation land models developed by different modeling groups, documenting the approaches used to represent the storage and transmission of water in soils, infiltration, root water uptake, groundwater dynamics, stream-aquifer interactions, and channel/floodplain routing. Appendix A reviews the similarities and differences among these land models, and contrasts the representation of hydrologic processes in land models with the modeling approaches used in the hydrologic sciences. The main inter-model differences in Table 2 are (1) the storage and transmission of water in soils is typically simulated using the moisture-form of Richards' equation; (2) infiltration is most commonly simulated as saturation-excess runoff, where saturated areas are parameterized either as a function of water table depth or soil moisture; (3) the lower boundary condition is often handled as free drainage or restricted drainage; (4) root water uptake is typically simulated using a macroscopic approach (i.e., ignoring hydraulic gradients between the roots and the soil); and (5) groundwater dynamics and stream-aquifer interactions are either neglected or simulated using overly simple methods.

Table 3 identifies candidate areas to improve the representation of hydrologic processes in land models. The key areas are (1) improve simulations of the storage and transmission of water in the soil matrix, obtained through (a) implementing the mixed form of Richards' equation [*Celia et al.*, 1990; *Maxwell and Miller*, 2005] and (b) explicitly representing macropore flow [*Beven and Germann*, 1982; *Weiler*, 2005; *Nimmo*, 2010; *Yu et al.*, 2014]; (2) improve representation of hydraulic gradients throughout the soil-plant-atmosphere continuum to improve simulations of root water uptake and evapotranspiration [*Baldocchi and Meyers*, 1998; *Mackay et al.*, 2003; *Bonan et al.*, 2014]; (3) improve representation of groundwater dynamics across a hierarchy of spatial scales, including improving "among grid" and "within grid" groundwater representations [*Famiglietti and Wood*, 1994; *Troch et al.*, 2003; *Miguez-Macho et al.*, 2007]; and (4) improve simulations of streamflow, by explicitly representing stream-aquifer interactions and improving parameterizations of channel/floodplain routing [*Qu and Duffy*, 2007; *Shen and Phanikumar*, 2010; *Miguez-Macho and Fan*, 2012a; *Pappenberger et al.*, 2012]. Underpinning all of these areas is the need to improve data sets on geophysical attributes, especially data on bedrock depth and permeability [*Tesfa et al.*, 2009; *Fan et al.*, 2015] and data sets on the physical characteristics of rivers [*Getirana et al.*, 2013; *Mersel et al.*, 2013; *Gleason and Smith*, 2014].

The philosophy underlying the recommendations in Table 3 is the need to explicitly resolve dominant hydrologic processes. This approach differs from the approach often adopted in the hydrologic modeling community where the preference is to lump multiple processes (e.g., use a parsimonious catchment-average storage-discharge functions), and implicitly represent the aggregate impact of unresolved processes [e.g., *Perrin et al.*, 2003; *Fenicia et al.*, 2014]. We push for a mechanistic approach, even where the

Table 2. Representation of Hydrologic Processes in Different Land Models

Model	Storage/Transmission of Water Through Soils	Infiltration	Lower Boundary Condition	Root Water Uptake	Groundwater Dynamics	Routing
CABLE	Richards' equation (Clapp Hornberger functions)	Unlimited	Restricted drainage	Macroscopic approach	None	None
Catchment	Empirical approximation to 1-D Richards' equation	Saturation-excess runoff, $F_{sat}=f(z_{wt})$	Two-way flux between the root zone and water table	Macroscopic approach	Base flow computed using TOPMODEL concepts	None
CLM	Richards' equation (Clapp Hornberger functions)	Saturation-excess runoff, $F_{sat}=f(z_{wt})$	Free drainage	Explicitly represent the hydraulic gradients[a]	Base flow computed using TOPMODEL concepts	Linear storage-outflow relationship
TESSEL	Richards' equation (van Genuchten functions)	Saturation-excess runoff, $F_{sat}=f(\theta)$	Free drainage	Macroscopic approach	None	1-D diffusive wave
JULES	Richards' equation (either Clapp Hornberger or van Genuchten functions)	Saturation-excess runoff; options for $F_{sat}=f(\theta)$ or $F_{sat}=f(z_{wt})$	Options for free drainage and zero-flux boundary conditions (depends on groundwater dynamics)	Macroscopic approach	Includes the option to use TOPMODEL concepts to compute base flow	None
LEAF	Richards' equation (Clapp Hornberger functions)	Infiltration-excess and Saturation-excess runoff, $F_{sat}=f(\theta)$	Water table represented as a moving lower boundary	Macroscopic approach	Grid-to-grid lateral flow based on Darcy's Law	1-D diffusive wave
LM3	Richards' equation (Clapp Hornberger functions)[b]	Unlimited[c]	Zero flux	Explicitly represent the hydraulic gradients	Hillslope flow based on Darcy's Law	1-D diffusive wave in low slope reaches
MATSIRO	Richards' equation (Clapp Hornberger functions)	Saturation-excess runoff, $F_{sat}=f(z_{wt})$	Water table represented as a moving lower boundary	Macroscopic approach	Base flow computed using TOPMODEL concepts	None
Noah	Richards' equation (Clapp Hornberger functions)	Based on variability in infiltration and precipitation	Restricted drainage	Macroscopic approach	None	None
ORCHIDEE	Richards' equation (van Genuchten functions)	Saturation-excess runoff, $F_{sat}=f(\theta)$	Free drainage[d]	Macroscopic approach	None	Linear storage-outflow relationship
VIC	Richards' equation (Clapp Hornberger functions)	Saturation-excess runoff, $F_{sat}=f(\theta)$	Free drainage	Macroscopic approach	Conceptual bucket below the soil profile	Source-to-sink method

[a]The representation of hydraulic gradients throughout the soil-plant-atmosphere continuum [*Bonan et al*., 2014] is not yet incorporated in an official release of CLM.
[b]LM3 has separate micropore and macropore domains; Richards' equation used for the micropore domain and water moves instantaneously through macropore domain.
[c]LM3 computes surface runoff if the infiltration capacity of the soil exceeded – however, the would-be surface runoff enters macropores (if macropores are present).
[d]*Campoy et al*. [2013] also experimented with options for (i) restricted drainage and (ii) imposing water table boundary condition at specific depths within the soil profile.

results compared to today's observations do not necessarily agree as well, because we cannot guarantee that the parsimonious lumped models will hold under a different climate regime [*Vaze et al*., 2010; *Milly and Dunne*, 2011].

An important issue is defining the appropriate balance between model complexity and model fidelity. We define model complexity as the number of model state variables and fluxes—in this definition model complexity describes the extent to which physical processes are represented explicitly as well as the discretization and connectivity of the physical landscape. We define model fidelity as the extent to which models faithfully represent observed processes. The trade-offs between complexity and fidelity bring up two central questions: first, to what extent is additional model complexity supported by the available information on geophysical attributes (topography, vegetation, soils, geology, and fine-scale meteorological data)? Or, put differently, does it make sense to include additional processes if there is limited information on the spatial variability of model parameters? Second, to what extent does the higher computational cost of more complex models limit their use, especially by constraining the application of continental-domain parameter estimation methods [*Samaniego et al*., 2010] and stochastic simulations [*Maxwell and Kollet*, 2008a]. We hence

Table 3. Opportunities for Future Land Surface Model Development

Model Development Opportunities	Expected Impact	References
Explicitly represent variably saturated flow using the mixed form of Richards' equation	Improve simulations of shallow groundwater dynamics and soil moisture	*Celia et al.* [1990], *Maxwell and Miller* [2005], and *Kumar et al.* [2009]
Explicitly represent vapor flow through soil	Improve simulations of evapotranspiration	*Parker et al.* [1987], *Painter* [2011], *Zeng et al.* [2011], and *Smits et al.* [2012]
Explicitly represent macropore and fracture flow	Improve simulations of soil moisture, evapotranspiration, groundwater dynamics, and runoff	*Beven* and Germann [1981, 1982], *SimunEk et al.* [2003], *Weiler* [2005], *McDonnell et al.* [2007], *Maxwell and Kollet* [2008a], *Maxwell*, [2010], *Nimmo* [2010], and *Yu et al.* [2014]
Explicitly represent reinfiltration of surface runoff as water moves across the landscape	Improve simulations of soil moisture and partitioning of precipitation into evapotranspiration and runoff	*Downer and Ogden* [2004], *Kollet and Maxwell* [2006], *Kumar et al.* [2009], and *Therrien et al.* [2010]
Explicitly represent hydraulic gradients throughout the soil-plant-atmosphere continuum	Improve simulations of root water uptake and evapotranspiration	*Baldocchi and Meyers* [1998], *Lai and Katul* [2000], *Williams et al.* [2001], *Mackay et al.* [2003], *Bonan et al.* [2014], and *Manoli et al.* [2014]
Explicitly represent "among-grid" groundwater flow, using 2-D or 3-D models	Improve simulations of groundwater dynamics and evapotranspiration	Miguez-Macho *et al.* [2007]. *Kumar et al.* [2009], *Therrien et al.* [2010], *Miguez-Macho and Fan* [2012a,b], and *Maxwell et al.* [2015]
Explicitly (or implicitly) represent "within-grid" groundwater flow, using representative hillslopes	Improve simulations of groundwater dynamics and evapotranspiration	*Famiglietti and Wood* [1994], *Wigmosta and Lettenmaier* [1999], *Beven and Freer* [2001], *Paniconi et al.* [2003], *Troch et al.* [2003b], and *Subin et al.* [2014]
Explicitly represent stream-aquifer interactions	Improve simulations of groundwater dynamics and streamflow	*Panday and Huyakorn* [2004], *Kollet and Maxwell* [2006], *Qu and Duffy* [2007], *Shen and Phanikumar* [2010], *Miguez-Macho and Fan* [2012a], and *Niu et al.* [2014]
Improve simulations channel/floodplain routing by implementing 1-D diffusive wave models	Improve simulations of streamflow, especially backwater effects	*Yamazaki et al.* [2011], *Miguez-Macho and Fan* [2012a], *Pappenberger et al.* [2012], and *Yamazaki et al.* [2012]
Improve data sets on bedrock depth and bedrock permeability	Improve simulations of soil moisture and groundwater dynamics	*Tesfa et al.* [2009] and *Fan et al.* [2015]
Improve data sets on physical characteristics of rivers (e.g., slope, roughness, hydraulic geometry)	Improve simulations of streamflow and stream-aquifer interactions	*Getirana et al.* [2013], *Mersel et al.* [2013], and *Gleason and Smith* [2014]

propose careful analysis of the trade-offs between complexity and fidelity; we will return to this issue in section 4.3.

3.2. Heterogeneity and Scaling Behavior

A critical issue in land modeling is to understand and reliably simulate the impacts of heterogeneity in interlinked hydrologic-biophysical-biogeochemical processes on land-atmosphere fluxes at larger spatial scales [*Giorgi and Avissar*, 1997; *Essery et al.*, 2003]. Model experiments show that heterogeneity in soil moisture can have profound impacts on regional meteorology [*Avissar and Pielke*, 1989; *Chen and Avissar*, 1994; *Mahrt*, 2000; *Huang and Margulis*, 2009], signifying that improved representation of the heterogeneity in soil moisture (and associated vegetation) can substantially improve simulations of land-atmosphere interactions [*Maxwell and Kollet*, 2008b].

The crux of the problem is to simulate how subgrid-scale heterogeneities are manifest in grid-average fluxes. This has received substantial attention in the land modeling community over the past several decades [*Shuttleworth*, 1988; *Entekhabi and Eagleson*, 1989; *Pitman et al.*, 1990; *Dolman and Gregory*, 1992; *Koster and Suarez*, 1992], and has been addressed in five main ways: (1) explicit representation of subgrid variability—this can be accomplished by configuring the land model with a finer mesh than the rest of the ESM [*Hahmann and Dickinson*, 2001], by using multiple tiles to represent subgrid heterogeneity [*Koster and Suarez*, 1992; *Bonan et al.*, 2002], or by explicitly representing the spatial variability for a subset of processes; e.g., separate stomatal conductance calculations for sunlit and shaded leaves [*Wang and Leuning*, 1998] or separate energy balance calculations for snow-covered and snow-free surfaces [*Takata et al.*, 2003; *Swenson and Lawrence*, 2012]; (2) statistical-dynamical models, which parameterize how subgrid variability in model state variables affects grid-average fluxes—for example, as discussed in section A1.1, the Probability Distributed Model [*Moore and Clarke*, 1981] and TOPMODEL [*Beven and Kirkby*, 1979] represent the impacts of subgrid variability in soil moisture and water table depth on grid-average infiltration [e.g., *Clark and Gedney*, 2008]; (3) development of new parameterizations that represent the aggregate impact of subgrid heterogeneities on grid-average fluxes [e.g., *Mahrt*, 1987; *Essery et al*., 2008]; (4) application of reduced-order modeling to construct surrogate models for subgrid variability from fine-grid simulations [e.g., *Pau et al*., 2014; *Riley and Shen*, 2014]; and (5) development of "effective" parameter values through application of upscaling operators that are specifically designed to reflect the impact of fine-scale heterogeneity on large-scale fluxes [e.g., *Samaniego et al*., 2010].

A key missing link in the current generation of land models is representing how the spatial organization of soil moisture and groundwater [*Western et al*., 1999; *Grant et al*., 2004] affects land-atmosphere fluxes [*Maxwell and Kollet*, 2008b]. In particular, most of the land models reviewed in Table 2 have a simplistic representation of the topographic controls on fine-scale soil moisture heterogeneity and the associated heterogeneity in evapotranspiration. Many models use a variant of the mosaic approach to represent surface heterogeneity, where land-atmosphere fluxes are computed separately for different vegetation types within a model grid box [*Koster and Suarez*, 1992]. A variant of the mosaic approach is used in CLM [*Bonan et al*., 2002], TESSEL [*Balsamo et al*., 2009], LEAF [*Walko et al*., 2000], VIC [*Liang et al*., 1994], and ORCHIDEE [*Campoy et al*., 2013]. In other models, the representation of spatial heterogeneity is even more limited—the Noah model [*Ek et al*., 2003], MATSIRO [*Takata et al*., 2003], and CABLE [*Kowalczyk et al*., 2006] have a single vegetation type per grid box; spatial heterogeneity is represented in CABLE solely through separate stomatal resistance calculations for sunlit and shaded leaves [*Dai et al*., 2004] and in MATSIRO through different surface energy calculations for snow covered and snow-free surfaces [*Takata et al*., 2003]. Two standout models with more detailed representations of subgrid heterogeneity are the Catchment model [*Koster et al*., 2000] and the tiled hydrology implementation of the LM3 model [*Subin et al*., 2014]—the Catchment model explicitly disaggregates a grid cell into catchments, with further disaggregation into zones of variable evapotranspiration stress; the LM3 model explicitly calculates the lateral flux of water among a sequence of tiles in a subgrid hillslope, leading to higher water table (and consequently lower plant water stress) in areas of topographic convergence.

There are a number of opportunities to improve the representation of the topographic controls on subgrid heterogeneities. Specifically, several existing approaches can readily be implemented into land models: (1) the distributed TOPMODEL approach, differing from the lumped TOPMODEL approach implemented in many land models (Table 2)—in the distributed approach, grid cells are disaggregated into multiple tiles based on the topographic index, with the local water table for a given topographic index class imposed as a lower boundary condition for each subgrid tile [*Famiglietti and Wood*, 1994]; (2) Kinematic subsurface flow along a representative hillslope—assuming that the water table gradient can be approximated by the topographic gradient, the groundwater flow equation is reduced to a 1-D kinematic wave equation and can be applied to simulate the water table position for various tiles along the hillslope [*Wigmosta and Lettenmaier*, 1999; *Beven and Freer*, 2001]; (3) dynamic subsurface flow across a representative hillslope, where lateral flux depends on spatial water table gradients—this is the approach used in the Hillslope Boussinesq model [*Troch et al*., 2003; *Paniconi et al*., 2003], and can be applied to multiple hillslope tiles to represent spatially variable evapotranspiration stress as in the LM3-TiHy model [*Subin et al*., 2014]; and (4) a high-resolution spatial mesh of subgrid tiles [*Hahmann and Dickinson*, 2001], which can be extended to explicitly resolve hill-to-valley convergence, implementing the same large-scale, among-grid groundwater flow model to finer grids within an ESM grid box. All of these methods represent both the spatial variability in physical characteristics (meteorology, vegetation, soils) as well as hydrologic connectivity, which enables simulations of the spatial organization of soil moisture and groundwater and plant water stress.

An important consideration is multiscale modeling of energy and mass fluxes, where different physical processes are simulated at different spatial resolutions. Multiscale modeling has a long history in land models: examples include separate stomatal conductance calculations for sunlit and shaded leaves, to scale fluxes from the leaf scale to the canopy scale [*Wang and Leuning*, 1998]; multiple plant functional types embedded within a single soil column to represent the impacts of spatial variability in vegetation phenology on grid-scale fluxes [*Bonan et al*., 2002]; and, more generally, a high-resolution spatial mesh of subgrid tiles to represent the large-scale impact of landscape heterogeneity [*Hahmann and Dickinson*, 2001]. More recent multiscale capabilities include high-resolution subgrid spatial mesh for terrain routing (subsurface and overland flow) and complex spatial representations of the river network and the built environment

[*Gochis et al*., 2013; *Voisin et al*., 2013a, 2013b; *Subin et al*., 2014; *Clark et al*., 2015a; *Li et al*., 2015]. The need for multiscale modeling capabilities introduces new software engineering requirements, including hierarchal data structures, unstructured grids, and more complex coupling procedures [*Oleson et al*., 2010; *Leonard and Duffy*, 2014].

3.3. Human Impacts on the Terrestrial Water Cycle

Human actions comprise another area ripe for improvement in ESMs. While humans are indirectly impacting the hydrologic cycle a myriad of ways including through anthropogenic climate change, e.g., through changes in snowpack hydrology [*Barnett et al*., 2005] and intensification of the water cycle [*Huntington*, 2006], humans are also impacting the water cycle directly. Construction of dams causes an increase of residence time on the land surface, fragments our waterways, and inundates river valleys [*Marble and Lough*, 1997]; surface water diversions distort our hydrologic regimes [*Poff et al*., 1997]; groundwater extraction and tile drains lower the water table with impacts on connected surface water bodies [*Zektser et al*., 2005]; irrigation mines surface and groundwater systems and increases evapotranspiration [*Boucher et al*., 2004]; and other land use changes such as forest and cropland management impact the partitioning of precipitation into ET, runoff, and groundwater recharge and flow [*Foley et al*., 2005].

Currently, many existing and new ESMs are being developed to resolve coupled human and natural systems, including representation of resource management activities [*Pokhrel et al*., 2012; *Adam et al*., 2014; *Kraucunas et al*., 2014]. CLM has been developed to represent maize, soybean, and spring wheat including management practices such as fertilizer application, residue management, and harvest [*Drewniak et al*., 2013]; irrigation from surface water [*Sacks et al*., 2009; *Leng et al*., 2013] and from groundwater [*Leng et al*., 2014]. VIC [*Liang et al*., 1994] has been fully coupled to a cropping system model (CropSyst), allowing for simulation of the interactions between large-scale hydrologic processes and crop growth, phenology, and management (including irrigation technology and management) [*Stöckle et al*., 2014]; and it has been adapted to simulate reservoir operations [*Haddeland et al*., 2006]. Other modeling frameworks that allow for representation of water management include WaterGap [*Alcamo et al*., 2003] the Model for Scale Adaptive River Transport (MOSART), which is a streamflow routing model [*Li et al*., 2013] that has been incorporated into an ESM framework [*Li et al*., 2015] and coupled with a water management model [*Voisin et al*., 2013a] and an integrated assessment model [*Voisin et al*., 2013b] to simulate water demand consistent with the socioeconomic scenarios. *Nazemi and Wheater* [2015a,b] provide a comprehensive review of the incorporation of water demand, supply, and allocation in ESM frameworks—they conclude that current models are limited in representing these processes, and that there is tremendous opportunity to improve models for understanding the role that water management plays in impacting earth system interactions.

3.4. Summary: Opportunities to Improve Representation of Hydrologic Processes in Land Models

The previous sections illustrate that the historical development of ESM land models can benefit from the advances in hydrologic research in the past decades. This is possible by improving the representation of key processes (e.g., groundwater dynamics), improving model representations of key aspects of spatial variability (e.g., bedrock depth, hydraulic geometry), improving model representations of the dominant features of the lateral flow of water above and below ground (e.g., reinfiltration of surface runoff, flow through macropores, river-aquifer interactions, and downstream controls on the flow of water through the river network), and improving model representations of the topographic controls on subgrid-scale heterogeneities in soil moisture and evapotranspiration fluxes.

To define a path forward, Table 3 summarizes key opportunities to improve the representation of hydrologic processes in land models. Active work on these topics is already underway in many modeling groups, and recent results prove that advances in the representation of hydrologic processes substantially improve the fidelity of simulations of land-atmosphere fluxes and biogeochemistry [e.g., *Miguez-Macho and Fan*, 2012a, 2012b; *Pappenberger et al*., 2012; *Milly et al*., 2014; *Subin et al*., 2014; *Swenson and Lawrence*, 2014]. Table 3 and the discussion in this section define what needs to be done—the next section outlines some strategic research tasks and considerations on how to improve hydrologic processes in ESMs.

4. Hydrologic Benchmarking Activities to Accelerate Modeling Advances

The opportunities to improve land models listed in Table 3 are gleaned from a diverse set of model development activities, making it difficult to understand the relative merits of competing modeling alternatives.

Model intercomparison projects have typically not yielded specific insight into the causes of differences in output across models because the extensive structural differences among the models make it difficult to attribute inter-model differences to specific modeling decisions [*Koster and Milly*, 1997; *Nijssen et al.*, 2003; *Clark et al.*, 2011]. There is therefore a strong need for a systematic and controlled approach to model evaluation in order to both identify preferable modeling alternatives and understand model weaknesses and model development needs.

In our opinion, improving the representation of hydrologic processes in land models demands comprehensive hydrologic benchmarking activities in order to systematically evaluate competing modeling alternatives. This requires four main elements: (1) synthetic test problems; (2) process-based model evaluation, making extensive use of multivariate and multiscale data from research basins; (3) evaluating the interplay between predictive accuracy, computational efficiency, and transferability; and (4) understanding the information content in data and models, in order to improve how land models use the data that are available to them. We will discuss each of these elements in the following sections, building on the existing literature on benchmarking of hydrologic and land models [*Abramowitz et al.*, 2008; *Gupta et al.*, 2008; *Blyth et al.*, 2011; *Clark et al.*, 2011; *Abramowitz*, 2012; *Gupta et al.*, 2012; *Luo et al.*, 2012; *Maxwell et al.*, 2014; *Nearing and Gupta*, 2014].

4.1. Synthetic Test Problems

Synthetic test cases are used to evaluate the implementation of the model equations, including impacts of the numerical approximations. For example, synthetic test problems have been used to evaluate different numerical solutions of Richards' equation [*Celia et al.*, 1990; *Boone and Wetzel*, 1996; *Tocci et al.*, 1997; *Lee and Abriola*, 1999; *Vanderborght et al.*, 2005; *Miller et al.*, 2006], different approaches to represent cyrosuction processes during soil freezing [*Hansson et al.*, 2004; *Noh et al.*, 2011; *Painter*, 2011], different approaches to simulate lateral subsurface flow [*Wigmosta and Lettenmaier*, 1999; *Rupp and Selker*, 2005, 2006; *Bogaart et al.*, 2013; *Troch et al.*, 2013], and different numerical approximations of overland flow [*Parlange et al.*, 1981; *Govindaraju et al.*, 1990; *Mizumura*, 2006; *Mizumura and Ito*, 2010, 2011]. Such synthetic test problems are not widely used in the land surface modeling community—yet there is a need to incorporate these synthetic test problems in model test suites in order to provide a rudimentary test of the model implementation and continually check if changes to the model corrupt the most basic model capabilities.

The synthetic test problems described above primarily focus on individual processes (e.g., infiltration into initially dry soil [*Celia et al.*, 1990]), and there is a need to augment the simple synthetic test cases with numerical experiments that evaluate interlinked physical processes. This was recently done by *Maxwell et al.* [2014], who used five test problems to evaluate the differences among seven integrated hydrologic models. The test problems included (1) infiltration-excess runoff for an inclined hillslope (saturated hydraulic conductivity less than the rainfall rate); (2) saturation-excess runoff for an inclined hillslope (saturated hydraulic conductivity greater than the rainfall rate); (3) surface runoff over a tilted V-shaped catchment; (4) runoff generation on a heterogeneous 1-D slab; and (5) return flow for an inclined hillslope. The results of this experiment showed strong inter-model agreement, most pronounced for the first three test problems, with inter-model differences attributed to differences in the numerical approximation [*Maxwell et al.*, 2014]. It will be important to include these "more advanced" test problems in land surface model test suites as the representation of hydrologic processes in land models continues to improve.

4.2. Process-Based Model Evaluation

The synthetic test problems described in section 4.1 are useful to evaluate the implementation of the model equations, but do not provide much insight on the more pressing need to evaluate the capability of different modeling approaches in real catchments. Synthetic test problems only provide the first (and the most basic) test of the model implementation. A systematic multivariate and multiscale evaluation of hydrologic modeling advances, making extensive use of data from research basins, is necessary to improve model representation of dominant processes [*Gupta et al.*, 2008; *Clark et al.*, 2011].

The process-based approach to evaluation has two related elements. First, it is important to minimize the number of differences between alternative model configurations so that it is possible to attribute differences in model behavior to individual modeling decisions. For example, in addition to comparing different channel routing models, it is important to evaluate the approximations/simplifications within a routing model. This can be accomplished using the so-called "multiple hypothesis" approach to model evaluation [*Clark et al*., 2011; *Beven*, 2012], through experimenting with different physics options, spatial configurations, and model parameter values, all within the same land surface model framework [*Best et al*., 2011; *Niu et al*., 2011; *Clark et al*., 2015a, 2015b]. The second element of process-based model evaluation is to pull together multivariate and multiscale observations to diagnose specific reasons for model inadequacies and guide future model development efforts [*Gupta et al*., 2008]. Multivariate and multiscale diagnostic metrics of model behavior should provide insight into both the individual processes within a model as well as how interlinked processes combine to produce the system-scale response at larger spatial scales [*Clark et al*., 2011; *Luo et al*., 2012]. For example, analysis of recession behavior can provide more information on model weaknesses than analysis of summary metrics of model performance such as the Nash-Sutcliffe score [*Yilmaz et al*., 2008; *Clark et al*., 2009]. Essentially, models need to reproduce both realistic temporal hydrologic behavior (i.e., high frequency and low frequency) and realistic spatial patterns of variability across a hierarchy of scales. As such, the process-based approach to model evaluation involves decomposing land models into a set of testable components (constituent hypotheses), and using multivariate and multiscale data to systematically evaluate individual model hypotheses and their interactions.

There are two main challenges in process-based evaluation of hydrologic modeling advances. The first challenge is a practical one: since land models must simulate processes in a myriad of different environments throughout the globe, there is a need to pull together the multivariate and multiscale data from a wide range of hydroclimatic regimes [*Blyth et al*., 2011]. The research data must be compiled in a consistent format, be serially complete (at least for the model forcing data), and include estimates of the uncertainty and spatial representativeness of individual observations [*Abramowitz*, 2012]. Many of the emerging data papers from research basins cover these issues to varying degrees [e.g., *Reba et al*., 2011; *Landry et al*., 2014], but there is much more work to be done to pull together research data from a wide range of watersheds.

The second challenge is to devise meaningful ways to use research data for model evaluation, accounting for problems of data uncertainty and the incommensurability of models and measurements [*Beven*, 1993; *Kuczera and Mroczkowski*, 1998; *Beven*, 2001; *Freer et al*., 2004; *Beven*, 2007; *Kuczera et al*., 2010], and making use of qualitative insights from experimentalists to constrain model behavior [*Seibert and McDonnell*, 2002]. A key modeling problem is representing how interlinked physical processes affect the aggregate system-scale responses at larger spatial scales [*Beven and Cloke*, 2012; *Wood et al*., 2012], and thus a key focal area for model evaluation is understanding the merits of different methods to represent spatial variability and hydrologic connectivity. Focus on these scaling issues will help identify the modeling approaches that can reliably represent the impacts of subgrid-scale heterogeneities on grid-average fluxes (see section 3.2). In a general sense, this second challenge can be addressed by pursuing the systematic and controlled approach to model evaluation detailed earlier [*Clark et al*., 2011, 2015a, 2015b]; however, the specific details on defining meaningful diagnostic signatures, the extent to which it is possible to isolate individual modeling decisions, and designing/implementing methods to characterize model and data uncertainty, are all important issues that require substantial effort and debate [*Beven et al*., 2012; *Clark et al*., 2012].

The challenge of multivariate and multiscale model evaluation will require a substantial amount of determination. Concerted effort on model evaluation, especially systemizing the model evaluation approach, will help improve the representation of hydrologic processes in land models, will help understand differences in performance across models and across model versions, and enable assessment of how improved hydrologic process representation affects other aspects of the land model system.

4.3. Interplay Between Predictive Accuracy, Computational Efficiency, and Data Availability

A key issue to consider as part of model evaluation is the trade-off between predictive accuracy (or model fidelity) and computational efficiency. This is of critical importance as land models begin to include a bewilderingly large set of processes [*Lawrence et al*., 2011]. For example, there is currently limited understanding of the cost-benefit trade-off of moving from single-layer canopy models to multilayer canopy models [*Bonan et al*., 2014], the shift from noniterative to iterative solutions of Richards' equation [*Zeng and Decker*, 2009; *De Rooij*, 2010; *Zeng and Decker*, 2010] and the importance of kinematic versus diffusive models to simulate lateral subsurface flow [*Beven and Freer*, 2001; *Paniconi et al*., 2003; *Troch et al*., 2003]. The lack of understanding of accuracy-efficiency trade-offs is especially acute in the land modeling community, as there is very little published work on how different numerical approximations of hydrologic processes affect the representation of land-atmosphere fluxes.

Another major issue is the tension between global applicability and data availability. For any parameterization to be implemented in an ESM, it has to be applicable globally (e.g., for all climate regimes, for all land cover types). For instance, while insights can be gained from detailed studies over specific river basins, a substantial amount of additional work is needed to extend such insights to all river basins (e.g., from tropical to polar river basins). To this end, effort is required to estimate important geophysical attributes on the global scale (e.g., soil information, bedrock depth, hydraulic geometry) [*Tesfa et al*., 2009; *Chaney et al*., 2014; *Fan et al*., 2015; *Allen and Pavelsky*, 2015], and use the available geophysical information to estimate spatially variable model parameters [*Samaniego et al*., 2010]. The broad applicability of modeling advances also requires a synthesis across catchment-scale observatories, to pull out processes and interactions that are common in a wide range of climatic-hydrologic-ecosystem regimes [*McDonnell et al*., 2007; *Wagener et al*., 2007]. As such, model benchmarking requires understanding the interplay between predictive accuracy, computational efficiency, and the availability of data to support competing modeling approaches.

4.4. Understanding the Information Content in Data and Models

Building on the synthetic test cases and process-based model evaluation, there is a need to advance the "science" of benchmarking through improving understanding of the information content in data and models. This issue relates to our expectations of model performance. Specifically, we ask the following related questions: how do we define a "good" model, and can we define meaningful benchmarks to evaluate the extent to which the model contains useful information [*Kirchner et al*., 1996; *Schaefli and Gupta*, 2007; *Abramowitz et al*., 2008; *Blyth et al*., 2011; *Gupta et al*., 2012; *Nearing and Gupta*, 2014].

The benchmarking problem involves comparing a proposed model to a baseline alternative (i.e., the null hypothesis), and evaluating if the proposed model adds information [*Nearing and Gupta*, 2014]. For example, *Abramowitz et al*. [2008] compare the performance of land models to statistical models based on empirical relationships between model forcing data and model fluxes, effectively comparing the information in the land surface model to the information content in the model forcing data. This study actually showed that the statistical models consistently outperformed the state-of-the-art land models, suggesting that land models underutilize the information in the meteorological forcing data [*Abramowitz et al*., 2008].

The value of these formal benchmarks is that they force us to take a step back from the traditional model development activities and ask, "What went wrong?" Or, more specifically, to ask why the complex models cannot make better use of the information that is available to them. One potential explanation is that the models have very strong a priori constraints—these constraints are evident in the choice of model parameter values, the choice of process parameterizations, the choice of spatial configurations, and the choice of the numerical schemes [*Mendoza et al*., 2015]. Addressing these issues requires careful and systematic scrutiny of the model representation of physical processes [*Gupta et al*., 2008; *Clark et al*., 2011], along with dialog between modelers and observationalists [*Seibert and McDonnell*, 2002], to improve the fidelity of model simulations.

5. Concluding Remarks

As *Water Resources Research* reaches its half century milestone, we look back on the immense accumulation of knowledge by the hydrologic sciences community through decades of observations and synthesis, and as we look forward in anticipation of global-scale environmental change with great uncertainties, we ask ourselves: what are the major gaps in our understanding of large-scale fluxes of water, energy, and biogeochemical cycles, and how can we contribute this knowledge toward advancing Earth System Models (ESMs), that are arguably the only quantitative tools available to foresee possible futures?

In this contribution to the *Water Resources Research* 50 year special issue, we take the opportunity to reflect on these questions. Through examples of hydrologic stores and fluxes, we contrast how hydrologists understand them and represent them in catchment-scale models versus how they are currently described in the large-scale land models in ESMs. This paper is not meant to be a comprehensive review of all aspects of terrestrial hydrology, but, rather, to identify opportunities to bridge between the catchment-scale focus in the hydrologic sciences and the global-scale focus in ESMs, and to propose some of the many paths forward that can be taken to incorporate the well-understood and large-scale hydrologic processes into the next generation of ESMs.

The examples presented in this paper point to the following general perspectives:

1. ESM land model development can benefit from recent advances in hydrology, both through incorporating key processes (e.g., groundwater-surface interactions) and new approaches to describe multiscale spatial variability and hydrologic connectivity;
2. Accelerating modeling advances requires comprehensive hydrologic benchmarking activities, in order to systematically evaluate competing modeling alternatives, understand model weaknesses, and prioritize model development needs; and
3. A stronger collaboration is needed between the hydrology and ESM modeling community, both through greater engagement of hydrologists in ESM land model development, and through rigorous evaluation of ESM hydrology performance in research watersheds or Critical Zone Observatories. Such coordinated efforts in advancing hydrology in ESMs have the potential to substantially impact energy, carbon, and nutrient cycle prediction capabilities through the fundamental role that hydrologic processes play in regulating these cycles.

We hope that the examples in this paper will motivate a series of detailed synthesis activities in the near future among the hydrologic and ESM communities, in order to create a comprehensive road map that will help accelerate advances in the representation of hydrologic processes in land models.

We conclude with a highly relevant and prescient quotation from a seminal paper by Peter Eagleson published in *Water Resources Research* in 1986 entitled "The emergence of global-scale hydrology" [*Eagleson*, 1986]:

> Emerging problems of environmental change and of long range hydrologic forecasting demand knowledge of the hydrologic cycle at global rather than catchment scale. Changes in atmosphere and/or landscape characteristics modify the earth's metabolism through changes in its biogeochemical cycles. The most basic of these is the water cycle which directly affects the global circulation of both atmosphere and ocean and hence is instrumental in shaping weather and climate. Defining the spatial extent of the environmental impact of a local land surface change, or identifying, for forecasting purposes, the location and nature of climatic anomalies that may be causally linked to local hydrologic persistencies requires global scale dynamic modeling of the coupled ocean-atmosphere-landsurface. Development, evaluation, verification, and use of these models requires the active participation of hydrologists along with a wide range of other earth scientists [*Eagleson*, 1978, page 6S]. . . He who controls the future of global-scale models controls the direction of hydrology [*Eagleson*, 1986, page 13S].

Appendix A: Process Representations in Land Models

A1. Storage and Transmission Through Soils

Improving simulations of the storage and transmission of water through soils is critical in order to improve the representation of the soil moisture-climate/weather feedbacks. While there are many facets to land-atmosphere interaction mechanisms, the basic framework is that soil moisture affects the partitioning of surface net radiation into sensible and latent heat flux (evaporative fraction), which consequently affects surface temperature, boundary layer properties, cloud formation, and in some cases the initiation of convection leading to precipitation [*Douville*, 2004; *Dirmeyer et al.*, 2006; *Koster et al.*, 2006, 2011; *Huang and Margulis*, 2013].

As documented in Table 2, most models use one-dimensional Richards' equation to simulate the time evolution of volumetric liquid water content [*Chen and Dudhia*, 2001; *de Rosnay et al.*, 2002; *Milly et al.*, 2014; *Clark et al.*, 2015b]

$$\frac{\partial \theta}{\partial t} = -\frac{\partial q}{\partial z} + S_{et} + S_{lf} \quad \text{(A1)}$$

where θ is the volumetric liquid water content, q is the vertical flux of liquid water (m s^{-1}), and S_{et} is the sink term for evapotranspiration (root water uptake) and S_{lf} is the source-sink term for the lateral flux of liquid water (s^{-1}). Note that in many models, $S_{lf}=0$. The vertical flux of liquid water q is formulated as a function of volumetric liquid water content (i.e., leading to the moisture-based form of Richards' equation)

$$q(z) = -K \frac{d\psi}{d\theta} \frac{\partial \theta}{\partial z} + K \quad \text{(A2)}$$

where $K=f(\theta)$ is the hydraulic conductivity (m s^{-1}) and $\psi=f(\theta)$ is the liquid water matric potential (m). The boundary conditions for equation (A1) are the fluxes at the soil surface (i.e., the infiltration of water into the soil profile, see section A1.1) and the fluxes at the bottom of the soil profile (section A1.2).

Ignoring differences in boundary conditions for the time being (see sections A1.1 and A1.2), the inter-model differences in Table 2 relate primarily to the details of the numerical implementation. The land models reviewed here use either the soil hydraulic functions of *Clapp and Hornberger* [1978] or *van Genuchten* [1980] for K and ψ (Table 2). Moreover, the different models use different numerical approximations in terms of the number of soil layers [*de Rosnay et al*., 2000], the splitting methods used to separate the diffusive and gravity terms in Richards' equation [*Kowalczyk et al*., 2006], the methods used to estimate diagnostic variables at layer interfaces [*Boone and Wetzel*, 1996; *Campoy et al*., 2013], the choice of noniterative and iterative solution method [*Chen and Dudhia*, 2001; *Takata et al*., 2003], and the ad hoc numerical fixes necessary in the moisture-based form of Richards' equation to account for supersaturated soil layers [*Boone and Wetzel*, 1996; *Chen and Dudhia*, 2001; *Yeh and Eltahir*, 2005a; *Best et al*., 2011]. Other inter-model differences relate to the representation of soil freezing processes, especially the representation of supercooled liquid water and the migration of liquid water toward the freezing front [*Luo et al*., 2003; *Niu and Yang*, 2006; *Swenson et al*., 2012]. The outlier models in Table 2 are the Catchment model which uses an empirical approximation to Richards' equation to calculate the time evolution of bulk root zone soil moisture [*Koster et al*., 2000; *Ducharne et al*., 2000], and CLM which subtracts the hydrostatic equilibrium soil moisture distribution from the vertical soil moisture profile before solving Richards' equation [*Zeng and Decker*, 2009; *De Rooij*, 2010; *Zeng and Decker*, 2010].

All of these land models, however, treat the storage and transmission of soil water significantly differently than many of the physically motivated models used in the hydrologic modeling communities, which include a more comprehensive set of processes including lateral flow, interflow, lowland depression storage, and exchange with groundwater. Moving forward, there is an opportunity to improve representation of storage and transmission of soil water in land models, while maintaining computational efficiency in mind, by (1) explicitly representing variably saturated flow, as is possible in the mixed form of Richards' equation [*Celia et al*., 1990; *Maxwell and Miller*, 2005; *Kumar et al*., 2009], to improve simulations of shallow groundwater dynamics; (2) explicitly represent airflow (vapor diffusion) through the soil to improve simulations of bare soil evaporation [*Parker et al*., 1987; *Painter*, 2011; *Zeng et al*., 2011; *Smits et al*., 2012]; and (3) explicitly represent macropore flow to simulate the nonuniform wetting of the soil matrix and the heterogeneity of flow paths at larger spatial scales [*Beven and Germann*, 1981, 1982; *SimunEk et al*., 2003; *Weiler*, 2005; *McDonnell et al*., 2007; *Maxwell and Kollet*, 2008a; *Nimmo*, 2010; *Yu et al*., 2014]. Some integrated models have incorporated these processes—for example, the coupling of ParFlow and CLM represents three-dimensional variably saturated flow and has now been applied at continental scales [*Maxwell et al*., 2015], CLM now parameterizes the diffusion of water vapor through a dry surface layer to improve simulations of bare soil evaporation [*Swenson and Lawrence*, 2014], and the LM3 model has a simple representation of macropores [*Milly et al*., 2014].

A1.1. Infiltration and Surface Runoff

The upper boundary condition in equation (A1), i.e., the surface infiltration, is given as

$$q(z=0) = P_x - e_s - q_s \quad \text{(A3)}$$

where P_x is the flux of water incident at the soil surface (kg m^2 s^{-1}), including throughfall and canopy drip when the soil surface is free of snow and drainage from the bottom of the snowpack when snow is present, e_s is the bare soil evaporation (kg m^2 s^{-1}), and q_s is the surface runoff (kg m^2 s^{-1}). Here we ignore canopy processes, snow processes, and evaporation processes (all of which are important), and focus on the differences in surface infiltration caused by differences in the parameterizations of surface runoff.

In contrast to their similarities in storage/transmission of water in soils, current land models include substantially different representations of surface runoff (Table 2) and therefore infiltration. Three main approaches are used:

$$q_s = 0 \tag{A4}$$

$$q_s = P_x F_{sat} + (1 - F_{sat}) \max(0, P_x - i_{\max}) \tag{A5}$$

$$q_s = \frac{P_x{}^2}{P_x + I_c} \tag{A6}$$

where $i_{\max}$ is the maximum infiltration rate (kg m^2 s^{-1}), typically parameterized as a function of volumetric liquid water content or liquid water matric potential in the upper soil layer(s), F_{sat} is the saturated fraction of the catchment, and I_c is the infiltration capacity (kg m^2 s^{-1}), parameterized as a function of the soil moisture deficit and the saturated hydraulic conductivity (see *Schaake et al.* [1996] and *Chen and Dudhia* [2001] for specific details on the parameterization of I_c). Specific details are as follows:

1. The first approach of $q_s = 0$ is used in only a small number of models, including CABLE [*Wang et al.*, 2011] and LM3 [*Milly et al.*, 2014]. In LM3, the flux at the upper boundary is initially computed as $q(z=0) = P_x - e_s$, with surface runoff computed if the solution to equation (A1) yields positive matric head (meaning that the infiltration capacity of the soil exceeded). However, the would-be surface runoff enters the soil if macropores are present and pore space is available, meaning $q_s = 0$ in many situations.
2. The second approach, defined in equation (A5), combines saturation-excess runoff and infiltration-excess runoff. This approach is used in most land models (Table 2), though in some models the infiltration-excess runoff—the second term in equation (A5)—is set to zero [e.g., *Liang et al.*, 1994; *Verant et al.*, 2004].
   a. Saturation-excess runoff depends on the parameterization of saturated fraction F_{sat}, where (i) F_{sat} is parameterized as a step function of soil moisture, with $F_{sat} = 1$ when the upper-most layer reaches capacity—this is the approach used in ORCHIDEE [*Verant et al.*, 2004]; (ii) F_{sat} is parameterized as a power law function of soil moisture, as used in VIC [*Liang et al.*, 1994], TESSEL [*Balsamo et al.*, 2009] and JULES [*Best et al.*, 2011]; and (iii) F_{sat} is parameterized as a function of the depth to the water table, as used in the Catchment model [*Koster et al.*, 2000], MATSIRO [*Takata et al.*, 2003], CLM [*Niu et al.*, 2007], and JULES [*Best et al.*, 2011]. Note that JULES has two options for saturation-excess runoff. The parameterization of the saturated fraction has a large impact on runoff dynamics—surface runoff can respond more rapidly to variations in soil moisture than to variations in water table depth, and associated differences in surface infiltration can affect base flow [*Boone et al.*, 2004; *Clark and Gedney*, 2008; *Clark et al.*, 2008].
   b. Infiltration excess depends on the parameterization of the maximum infiltration rate $i_{\max}$. In models where $i_{\max}$ is defined as the saturated hydraulic conductivity of the upper-most soil layer [e.g., *Takata et al.*, 2003] lead to the situation where $i_{\max}$ is typically much greater than P_x, meaning that infiltration-excess runoff is rather rare [*Balsamo et al.*, 2009].
3. The third approach, defined in equation (A6), is only used in the Noah model [*Ek et al.*, 2003], equation (A6) was formulated by *Schaake et al.* [1996] to represent the spatial variability in precipitation and infiltration capacity.

The similarity among models is their simplistic representation of the heterogeneity of flow paths. Runoff in most models is represented simply as the sum of surface runoff and base flow. Most models do not represent exfiltration, which is important to simulate return flow and wetland dynamics, and no allowance is made for the reinfiltration of surface runoff. Such simplifications can affect even the most basic hydrologic function such as the partitioning of precipitation among evapotranspiration and runoff.

A key issue here is identifying reliable methods to represent the impact of small-scale heterogeneity on the large-scale fluxes of infiltration and surface runoff, considering the merits of different modeling approaches to represent heterogeneity and scaling behavior (section 3.2) and the interplay between predictive accuracy, computational efficiency, and data availability (section 4.3). The approaches typically used in land models parameterize the impacts of subgrid-scale heterogeneities on the grid-average infiltration flux (Table 2), considering the aggregate impacts of subgrid variability in soil moisture [*Moore and Clarke*, 1981; *Wood et al*., 1992; *Clark and Gedney*, 2008; *Balsamo et al*., 2009], or the aggregate impacts of subgrid variability in water table depth [*Beven and Kirkby*, 1979; *Gedney and Cox*, 2003; *Niu et al*., 2005]. This is typically known as the top-down modeling approach [*Klemes*, 1983; *Dooge*, 1986]. By contrast, large-scale infiltration in physically based hydrologic models is represented as the spatial average across a high-resolution spatial mesh [*VanderKwaak and Loague*, 2001; *Downer and Ogden*, 2004; *Kollet and Maxwell*, 2006; *Qu and Duffy*, 2007; *Kumar et al*., 2009; *Therrien et al*., 2010]. This approach is typically known as the bottom-up approach to hydrologic modeling, and is elegantly presented in a blueprint for distributed hydrologic modeling by *Freeze and Harlan* [1969]. The physically based models generally include adherence to continuity in flux and head, which allows the possibility of flow in either direction (i.e., infiltration and exfiltration). Physically based models also implement a more detailed representation of wetting front dynamics [e.g., *Garrote and Bras*, 1995; *Ivanov et al*., 2004], which may improve simulations of infiltration and surface runoff during heavy rainfall. The best path forward is not immediately apparent—for example, *Maxwell and Kollet* [2008a] illustrate some of the difficulties of applying the top-down approach to estimate the large-scale infiltration flux. There is clearly an opportunity to use the physically based methods to explicitly simulate exfiltration, overland flow, and reinfiltration of surface runoff, though in the foreseeable future it may be necessary to parameterize some of these processes in order to maintain computational efficiency and tractability. We discuss these issues in the benchmarking discussion (section 4).

A1.2. The Bottom Boundary Condition for Soil Hydrology

The bottom boundary condition in equation (A1) is given as

$$q(z=z_u)=\begin{cases} c_d K_i & \textit{flux boundary condition} \\ -K_{sat}\dfrac{-\psi(\theta_i)}{z_u-z_i}+K_{sat} & \textit{head boundary condition} \end{cases} \tag{A7}$$

where z_u is the depth of the unsaturated zone (m), which is the depth of the water table for the head boundary condition, z_i is the depth of the lowest soil layer (m), K_i is the hydraulic conductivity of the lowest soil layer (m s^{-1}), K_{sat} is the saturated hydraulic conductivity (m s^{-1}), θ_i is the volumetric liquid water content of the lowest soil layer, and c_d is a parameter that controls the rate of flow at the lower boundary.

The key differences in the lower boundary condition are as follows:

1. The flux boundary condition is used in most land models (Table 2), but with important differences in the specification of the c_d parameter. The different approaches include (a) a zero-flux lower boundary, i.e., $c_d=0$, used in LM3 [*Milly et al*., 2014]; (b) a free drainage lower boundary, i.e., $c_d=1$, used in VIC [*Liang et al*., 1996], CLM [*Oleson et al*., 2010], ORCHIDEE [*Campoy et al*., 2013], and TESSEL [*ECMWF*, 2014]; and (c) restricted vertical drainage, i.e., $c_d < 1$ as used in the Noah model [*Ek et al*., 2003] and CABLE [*Wang et al*., 2011].
2. The head boundary condition, i.e., moving lower boundary defined by the depth of the water table, is now used in LEAF [*Miguez-Macho et al*., 2007] and MATSIRO [*Koirala et al*., 2014]. This is an effective way of separating the "unsaturated" zone from the "saturated" zone. However, the flux parameterization in equation (A2), i.e., the moisture-based form of Richards' equation, as used in LEAF and MATSIRO, can only simulate unsaturated flow and this separation of the saturated and unsaturated zones complicates representing shallow groundwater dynamics such as perched water tables. A variant of this boundary condition, used in PAWS+CLM [*Shen and Phanikumar*, 2010; *Shen et al*., 2013], is a noniterative coupling scheme between saturated groundwater flow and the unsaturated zone. With this method, the water table simulated by the saturated zone is consistent with the soil moisture profile. This class of methods can dynamically describe return flow, regional groundwater circulation, groundwater discharge, and saturation excess runoff [see also *Camporese et al*., 2009; *Camporese et al*., 2010].

The choice of the bottom boundary condition, along with the depth of the unsaturated zone, can have a large impact on the vertical soil moisture profile and land-atmosphere fluxes [e.g., *Zeng and Decker*, 2009; *Gochis et al*., 2010; *Campoy et al*., 2013]. The free drainage lower boundary condition that is still used in many models assumes spatially uniform soil depth [*Ek et al*., 2003] and "no bedrock anywhere" [*ECMWF*, 2014]. These representations do not consider the upward capillary flow or the vertical fluxes arising from negative potentiometric gradient due to evapotranspiration-induced dryness in the vadose zone. The free drainage lower boundary condition hence effectively neglects the interactions between surface water and groundwater that can be key drivers of land-atmosphere fluxes.

Improvements in land model simulations of soil moisture dynamics are possible through improving simulations of groundwater dynamics (see section A3), which involves more sophisticated representation of the flux at the bottom of the vadose zone. This will require spatially variable datasets of bedrock depth [*Tesfa et al.*, 2009] as well as datasets of permeability and deeper sediments below the depth covered by soil data [*Fan et al.*, 2015].

A2. Root Water Uptake

Root water uptake, represented by the evapotranspiration sink term S_{et} in equation (A1), often represents a large flux of water from the subsurface, and as such has long been recognized as an important term in hydrologic models [e.g., *Feddes et al.*, 1974]. Substantial evidence has accumulated that roots can be adaptive and dynamic [e.g., *Harper et al.*, 1991] and that roots can be shaped by local variations in soil water, oxygen, and nutrient status [e.g., *Hodge*, 2004]. Additionally, the presence of a shallow groundwater may profoundly influence root structure and dynamics, the latter in turn shaping land hydrology and regulating water, energy, and biogeochemical pools and fluxes [*Norby and Jackson*, 2000]. This new body of knowledge is directly relevant to understanding ecosystem response-feedback to a rapidly changing global environment, and to improving global model capabilities to predict the possible trajectories of the coevolution of the ecosystems and the physical world [*Fan*, 2015].

As documented in Table 2, root water uptake is represented in most land models using a macroscopic approach described by *Feddes et al.* [2001]. In this approach, the root water uptake for a given soil layer *j* is given as

$$S_{et,j}=\frac{f_{roots,j}\beta_{v,j}}{\bar{\beta}_v}\frac{Q^{veg}_{trans}}{L_{vap}\rho_{liq}\Delta z_j} \tag{A8}$$

where Q^{veg}_{trans} (W m^{-2}) is the transpiration latent-heat flux, $\beta_{v,j}$ is the soil water stress factor for the *j*th soil layer, $\bar{\beta}_v$ is the total water availability stress factor, $f_{roots,j}$ is the fraction of roots in the *j*th soil layer, Δz_j is the depth of the *j*th soil layer (m), and L_{vap} (J kg^{-1}) and ρ_{liq} (kg m^{-3}) are, respectively, the latent heat of vaporization and the intrinsic density of liquid water. The soil water stress factor in equation (A8) is typically parameterized as a function of volumetric liquid water content [e.g., *de Rosnay et al.*, 2002; *Best et al.*, 2011; *Wang et al.*, 2011]

$$\beta_{v,j}=\frac{\theta_j-\theta_{wlt}}{\theta_{crit}-\theta_{wlt}} \qquad \text{for } \theta_{wlt}<\theta_j<\theta_{crit} \tag{A9}$$

where θ_{crit} and θ_{wlt} define respectively the soil moisture content when plants become stressed and the residual soil moisture content (note $\beta_{v,j}=0$ for $\theta_j<\theta_{wlt}$ and $\beta_{v,j}=1$ for $\theta_j>\theta_{crit}$). The total water availability stress factor is

$$\bar{\beta}_v=\sum_j f_{roots,j}\beta_{v,j} \tag{A10}$$

which is used in the parameterizations of stomatal conductance. An important point here is that equation (A8) does not account for the hydraulic gradients between soil and roots.

The vertical root distribution is typically parameterized as a simple time-invariant (but spatially variable) function [*de Rosnay and Polcher*, 1998]

$$R(z)=e^{-cz} \tag{A11}$$

where $R(z)$ defines the fraction of roots below depth *z* and the parameter *c* is fit for different vegetation types. *Jackson et al.* [1996] compile a database of 250 different root studies to estimate the vertical root distribution for 11 biomes, and these vertical root distributions have been used to estimate rooting parameters for vegetation classes that are commonly used in land models [*Zeng et al.*, 1998; *Zeng*, 2001]. Variants of equation (A11) are used in CABLE [*Wang et al.*, 2011], CLM [*Oleson et al.*, 2010], TESSEL [*ECMWF*, 2014], and ORCHIDEE [*d'Orgeval et al.*, 2008].

The typical land model approach to simulate root water uptake as described here has several limitations. First, parameterizing the soil stress function based only on soil moisture does not represent the hydraulic gradient between soils and plants, which is a critical driver of evapotranspiration. Other modeling approaches explicitly simulate the hydraulic gradient throughout the soil-plant-atmosphere continuum [*Baldocchi and Meyers*, 1998; *Williams et al*., 2001; *Manoli et al*., 2014] and these approaches have recently been shown to provide substantial benefits in land models [*Hickler et al*., 2006; *Bonan et al*., 2014]. A limitation of this approach to date is that it is based on setting a minimum canopy xylem pressure, and so it can underpredict drought responses when scaling up to larger areas [*Powell et al*., 2013]. An alternative is to model the soil-plant hydraulic continuum using both the hydraulic properties of soil and cavitation in plant xylem represented with s-shaped curves representing the distinct changes in root, stem, and leaf hydraulic conductances with declining xylem pressure [*Sperry et al*., 1998]. This approach has been used to extend the TREES model [*Mackay et al*., 2003] to accurately predict root water uptake in response to drought [*McDowell et al*., 2013].

A second limitation is that the root profiles in the current generation of land models are time-invariant, which contrasts with the natural system where roots respond dynamically to changes in available energy and water [*Schenk and Jackson*, 2002]. Based on recent developments, there is an opportunity here to implement dynamic root growth models in land models, e.g., based on optimality principles [*Schymanski et al*., 2008]. Such models can utilize optimal rooting strategies following directly from soil-rhizosphere-plant hydraulic theory, which show that deeper roots are important for survival during drought [*Lai and Katul*, 2000], the coevolution of root to leaf area ratios ensure sufficient, yet not inefficient, water uptake by roots [*Ewers et al*., 2000; *Sperry et al*., 2002; *Zea-Cabrera et al*., 2006], and competition for water among roots of cooccurring plants [*Domec et al*., 2012; *Quijano et al*., 2012] to more directly link biophysical and hydrologic processes and improve simulations of root water uptake and evapotranspiration. Indeed, evidence of coordination between plant hydraulic traits on a global scale [*Manzoni et al*., 2013] should enable constraining future root water uptake models for use in land surface models.

A3. Groundwater Dynamics

There is increasing evidence that groundwater dynamics have an important impact on weather and climate [*Maxwell and Kollet*, 2008b]. A compilation of water table depth observations suggests that the groundwater is sufficiently shallow to influence the root-zone soil water over notable portions of the land area [*Fan et al*., 2013], and a data-model synthesis of groundwater-surface interactions in the Amazon suggests that groundwater plays an important role in regulating surface fluxes [*Miguez-Macho and Fan*, 2012a,b], and these processes may be underrepresented in most current land models

To date, five main approaches have been used to simulate groundwater dynamics:

1. No explicit representation of groundwater dynamics, as in the Noah model [*Ek et al*., 2003], CABLE [*Wang et al*., 2011], ORCHIDEE [*Koirala et al*., 2014], and TESSEL [*ECMWF*, 2014]. In this approach, runoff is simply the sum of surface runoff and drainage from the bottom of the soil profile [*Boone et al*., 2004].
2. A base flow bucket below the soil profile, as used in VIC [*Liang et al*., 1996]. In this approach, drainage from the soil profile flows into a "conceptual bucket" and nonlinear storage-discharge relationships are used to calculate base flow. There is no upward (capillary) flow from the bucket to the soil profile, hence the "saturated" storage has no impact on soil moisture dynamics. Similar approaches are used in the Total Runoff Integrating Pathways (TRIP) model [*Alkama et al*., 2010; *Decharme et al*., 2010; *Decharme et al*., 2012]. However, VIC-GW [*Liang et al*., 2003; *Leung et al*., 2011] allows for two-way soil-water table interactions in the vertical direction.
3. Application of general TOPMODEL concepts, where base flow from the soil profile is parameterized based on a statistical distribution of water table depth, as used in the Catchment model [*Koster et al*., 2000], JULES [*Gedney and Cox*, 2003; *Best et al*., 2011], MATSIRO [*Yeh and Eltahir*, 2005a,b; *Koirala et al*., 2014], and CLM [*Niu et al*., 2007; *Oleson et al*., 2010]. This is accomplished either by using a source-sink term in Richards' equation, i.e., the S_{lf} term in equation (A1) [*Gedney and Cox*, 2003; *Niu et al*., 2007], or by representing the water table as a moving lower boundary condition for the soil profile [*Koster et al*., 2000; *Yeh and Eltahir*, 2005a,b; *Koirala et al*., 2014].
4. The grid-to-grid (or column-to-column) lateral flux computed using Darcy's Law, as used in LEAF [*Miguez-Macho et al*., 2007; *Fan and Miguez-Macho*, 2011] and LM3 [*Milly et al*., 2014; *Subin et al*., 2014].
5. Coupling a complete integrated hydrology model with the biophysical components of a land model [*Maxwell and Miller*, 2005; *Maxwell et al*., 2011; *Shi et al*., 2013].

As noted above the *raison d'être* of land models is to provide the lower boundary condition for the atmosphere, so the interest here is the extent to which the different model implementations of groundwater affect soil moisture dynamics and land-atmosphere fluxes [*Maxwell et al*., 2007; *Maxwell and Kollet*, 2008b]. Interestingly, with the notable exception of the Catchment model [*Koster et al*., 2000], none of the TOPMODEL implementations represent the impact of subgrid variability in water table depth on soil moisture dynamics. Some more recent model implementations do explicitly represent the impact of spatial variability in water table depth, either through calculating the lateral flux from grid-to-grid [*Fan and Miguez-Macho*, 2011; *Miguez-Macho and Fan*, 2012a,b; *Shi et al*., 2014; *Maxwell et al*., 2015] or by calculating the lateral flux among a sequence of tiles in a subgrid hillslope [*Subin et al*., 2014].

The key challenge is to explicitly represent the impact of the spatial variability in water table depth on soil moisture dynamics and land-atmosphere fluxes across a hierarchy of spatial scales. At the current and projected near-future scales over which these land models will be applied, this necessarily requires simulating both large-scale "among-grid" and the smaller-scale "within-grid" groundwater dynamics. The large-scale "among-grid" dynamics can be represented using a 2-D (vertically integrated) groundwater flow equation (Dupuit-Forchheimer approximation) [*Fan and Miguez-Macho*, 2011; *Miguez-Macho and Fan*, 2012a,b] or full 3-D implementation of the mixed form of Richards' equation [*Maxwell et al*., 2015]. The smaller-scale "within-grid" dynamics can be represented by using the subgrid distribution of water table depth as the lower boundary condition for different subgrid tiles (e.g., where the different tiles are defined as areas with distinctive topographic convergence), or through extending the same large-scale, among-grid groundwater flow model to finer grids, without subgrid characterization at all (see section 3.2) The hillslope width function introduced by *Fan and Bras* [1998] is one example of a subgrid landscape configuration which imparts characteristic hydrologic responses in hillslope runoff formulations [*Troch et al*., 2003; *Hilberts et al*., 2004; *Bogaart and Troch*, 2006; *Carrillo et al*., 2011]. Additionally, variable resolution models, such as tRIBS [*Ivanov et al*., 2004; *Vivoni et al*., 2004; *Vivoni et al*., 2005] provide another means of representing different mechanisms of hydrologic complexity based on landform criteria, and such variable resolution approaches are currently being developed for broader earth system applications as well [*Skamarock et al*., 2012]. At this point, it is unclear which approach will be most effective, and systematic comparison of these different approaches will be necessary to identify the best path to improve representation of groundwater dynamics in ESMs. We expand on this point in the benchmarking discussion (section 4).

A4. Stream-Aquifer-Land Interactions

Only a subset of models reviewed here explicitly represents groundwater dynamics, and a much smaller subset of models explicitly represents two-way stream-aquifer interactions. One example of the latter is the LEAF model, which incorporates river elevation as part of the 2-D vertically, integrated groundwater flow equation, and allows river and floodwater to infiltrate through sediments in the flood plain [*Miguez-Macho et al*., 2007; *Miguez-Macho and Fan*, 2012a] that are an important source for groundwater recharge in many parts of the world [e.g., *Taylor et al*., 2013]. CLM also allows for reinfiltration of flooded waters, but without explicitly linking river dynamics with groundwater flow processes [*Oleson et al*., 2010].

It is well known in the groundwater modeling community that when groundwater is explicitly represented, streams and water bodies present critically important boundary conditions that strongly control the water table surface [*Harbaugh et al*., 2005]. However, most of the models reviewed here simulate groundwater dynamics using a kinematic approximation [*Liang et al*., 1994; *Koster et al*., 2000; *Oleson et al*., 2010; *Best et al*., 2011; *Koirala et al*., 2014], and hence only simulate one-way flow from the aquifer to the stream. The widespread disregard of stream-aquifer interactions in land models contrasts strongly with the modeling approaches used in integrated hydrologic models (e.g., ParFlow [*Kollet and Maxwell*, 2006], PIHM and FIHM [*Qu and Duffy*, 2007; *Kumar et al*., 2009], HydroGeoSphere [*Therrien et al*., 2010], and PAWS+CLM [*Shen et al*., 2013]).

Explicitly representing the two-way exchange of water between aquifers and rivers is critical in order to both provide boundary conditions to the groundwater flow system and to provide estimates of base flow, which is necessary to adequately simulate riparian ponding features (wetlands) and biogeochemical cycles. In the future, a more integrated representation of the terrestrial hydrologic cycle is desired. This requires the following modeling advances: (1) stream routing should move from source-to-sink schemes to reach-by-reach (or high resolution, gridded) routing schemes—the diffusive wave equation (section A.5) is a favorable approach due to its simplicity and stability; (2) the channel reaches need to interact with the land grid cells—stream-aquifer interaction is dynamic depending on flow stage, as well as the phase of the flood wave passage (i.e., whether or not the channel flow is rising or falling), requiring physically based representation or subgrid parameterization of groundwater head and river stage; and (3) important channel geometries, including channel width and bankfull depth, need to be provided or inferred from available datasets. While these conditions appear challenging, several hydrologic models have implemented these schemes [*Downer and Ogden*, 2004; *Qu and Duffy*, 2007; *Shen and Phanikumar*, 2010], and parsimonious representations of these processes have already been implemented in some land models [*Miguez-Macho et al.*, 2007; *Miguez-Macho and Fan*, 2012a,b].

A5. Channel/Floodplain Routing

The routing of water through channels and floodplains is a critical component of ESMs because it provides the freshwater flux to the oceans. Reliable simulations of channel/floodplain routing are also important for coupled simulations of groundwater dynamics (as described in section A4) and to enable detailed evaluation/improvement of the terrestrial hydrologic cycle in land models (see section 4). In many flat areas near-stem streams such as the Amazon, mid-stream Yangtze River, the Sudd, and the Okavango delta, channel-land exchange is a major control of the ecosystems of the floodplains and nearby regions [*Fan and Miguez-Macho*, 2011].

Three main approaches have been used for horizontal routing of runoff through river networks:

1. Source-to-sink methods, which use an impulse response function to calculate the travel time from individual grid boxes (sources) to a specific point on the river network (a sink) [*Lohmann et al.*, 1996; *Nijssen et al.*, 2001]. These methods therefore estimate the lumped response for all grid boxes above a desired sink (e.g., above a gauging station), assuming the runoff routing processes are linear and stationary [*Li et al.*, 2013]. The source-to-sink approach is used in the VIC model [*Nijssen et al.*, 2001], and has been applied as a postprocessing step for multiple land models [*Lohmann et al.*, 1998a, 2004; *Xia et al.*, 2012a, 2012b].
2. Empirical storage-outflow relationships, which calculate the flow in a given reach as a linear or nonlinear function of reach storage. Linear reach storage-outflow relationships are used in ORCHIDEE [*Ngo-Duc et al.*, 2007; *Guimberteau et al.*, 2012] and CLM [*Swenson et al.*, 2012], and nonlinear storage-outflow relationships are used in LM3 [*Milly et al.*, 2014]. For example, in CLM

$$Q_R = \upsilon \frac{S_R}{d} \tag{A12}$$

where Q_R is the outflow from a given reach (m s^{-1}), υ is the flow velocity (m s^{-1}), computed as a function of gridcell-average slope, S_R is the reach storage (m), and d is the stream distance (m), given as the distance among adjacent grid cells. These empirical storage-outflow relations are in the same class of models as the empirical Manning equation, implemented recently in CLM [*Li et al.*, 2013, 2015]

$$Q_R = \frac{\sqrt{\beta}}{n} k {R_h}^{2/3} \tag{A13}$$

where β is the channel slope, n is the Manning's roughness coefficient, k is a conversion factor, and R_h is the hydraulic radius (m), defined as the wetted area (m^2) divided by the wetted perimeter (m); with $R_h = f(S_R)$. Empirical storage-outflow methods assume that the gravity force dominates over all others [*Li et al.*, 2013].

3. The third approach used is 1-D diffusive wave models, in which water surface slope, instead of riverbed slope as in the kinematic wave method above, drives the flow, thus accounting for the backwater effect in flat river reaches. These models are used in TESSEL [*Yamazaki et al.*, 2011; *Pappenberger et al.*, 2012; *Yamazaki et al.*, 2012], LEAF [*Miguez-Macho and Fan*, 2012b], and LM3 [*Milly et al.*, 2014]. In LM3, the 1-D diffusive wave models are only used in reaches where the channel slope is below some specified threshold.

These 1-D channel routing models are more efficient and have lower data requirements than the 2-D and 3-D hydraulic models used in local-scale studies [*Wu*, 2008], and 1-D routing models are thus more suitable for use in global land models. In contrast to other hydrologic processes where developments in the hydrologic sciences have immediate applicability in land models, the global-scale applicability of complex hydraulic models is clearly rather limited. As such, we recommend further development and refinement of 1-D channel routing models in land models. In practice, this will mean shifting to models that explicitly represent the physical characteristics of rivers and floodplains [*Li et al.*, 2013]. This will require more widespread implementation of diffusive wave models in order to simulate downstream controls on flow (backwater effects) [*Miguez-Macho and Fan*, 2012a; *Pappenberger et al.*, 2012; *Yamazaki et al.*, 2013]. It may also require implementing a multiscale modeling approach within various hydrologic components of ESMs [*Ivanov et al.*, 2004; *Vivoni et al.*, 2004; *Vivoni et al.*, 2005; *Qu and Duffy*,2007; *Kumar et al.*, 2009] in order to resolve how complex landscape morphology affects hydraulic connectivity and intermittent channel flow.

A key model development need is to improve methods to estimate physical river characteristics (e.g., channel slope, roughness, and hydraulic geometry) and model parameters (e.g., the flow velocity parameter used in the simpler parameterizations). This is a challenge for all flow routing models—for example, the method to estimate the flow velocity parameter has recently been shown to be very important in linear storage-outflow models [*Swenson et al.*, 2012; *Li et al.*, 2015]. Many recent advances have gradually enabled us to address these challenges. River bed elevations extracted using a high-resolution DEM and a forward moving average filter in PAWS+CLM algorithm agree well with the groundwater table, permitting a highly efficient solver for the diffusive wave equation [*Shen et al.*, 2014]. Globally-available, 1/8° resolution river flow direction, accumulation and slope hydrography datasets, e.g., HydroSHEDS [*Lehner et al.*, 2008] and DRT [*Wu et al.*, 2012], have emerged. Remote-sensing data sets and the application of scaling laws provides global estimates of recharge [*Gleason and Smith*, 2014] and water depths [*Mersel et al.*, 2013]. The next generation remote-sensing products (e.g., the Surface Water and Ocean Topography, SWOT, expected to launch in 2020) will further expand our ability to characterize stream geometries and discharges [e.g., *Getirana et al.*, 2013; *Pavelsky et al.*, 2014; *Pedinotti et al.*, 2014; *Allen and Pavelsky*, 2015].

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