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Incorporating groundwater flow in land surface models: literature review and recommendations for further work

Groundwater Directorate

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Incorporating groundwater flow in land surface models: literature review and recommendations for further work

S Collins

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Summary

HydroJULES is a NERC-funded project that brings together NERC Centre-Surveys to investigate how to improve the simulation of the whole hydrological cycle in models. BGS' role is to inform the inclusion of groundwater in both the land surface model Joint UK Land Environment Simulation (JULES) (Best et al., 2011; Clark et al., 2011) and the hydrological model Grid to Grid (e.g. Bell at al, 2009). To facilitate this a literature review has been undertaken of the current methods for inclusion of groundwater in land surface models. The keywords 'global groundwater model', 'land surface models' and 'parameterisation'/'parameterization' were used to search the literature. Further, the main global datasets of relevance to HydroJULES have been summarised. The main finding is that the LEAF-Hydro approach (Miguez-Macho et al., 2007) is one of the most practical methods in the literature for including groundwater simulation in a land surface model. It is recommended that the LEAF-Hydro approach should be tested against existing BGS groundwater flow models for the UK.

1 Introduction

The land surface model Joint UK Land Environment Simulation (JULES) (Best et al., 2011; Clark et al., 2011) is a key component of NERC's Earth System Modelling Strategy. It is used in global and kilometre-scale weather forecasting, global climate prediction and earth system modelling. Despite being at the cutting edge of international land surface modelling, particularly with regard to mass and energy exchanges with the atmosphere, its soil–hydrological components are highly constrained. HydroJULES is a NERC-funded project that brings together NERC Centre-Surveys – specifically the Centre for Ecology and Hydrology, BGS and the National Centre for Atmospheric Science – to investigate how to improve the simulation of the whole hydrological cycle in models. BGS' role is to inform the inclusion of groundwater in both JULES and the hydrological model Grid to Grid (e.g. Bell at al, 2009). To facilitate this a literature review has been undertaken of the current methods for inclusion of groundwater in land surface models.

If we want to project the impacts of future climates, the way in which vegetation, soil and snow exchange water, energy and carbon with the atmosphere must be considered (Pitman, 2003). This is achieved with land surface models (LSMs), which provide physics-based descriptions of the processes involved.

Traditionally, LSMs focused on near-surface hydrology using the 1D Richards equation to calculate vertical flow in the soil (e.g. Gedney and Cox, 2003; Yeh and Eltahir, 2005a). The first LSMs did not include any simulation of groundwater, but instead applied a free drainage boundary condition to the bottom of a fixed soil column. However, in the last 20 years, the number of LSMs and their capabilities have increased significantly and groundwater simulation is now included in LSMs of all scales, from single basins to the globe. This report explores the range of methods used, from simplified lumped models (Section 2) to complex and computationally expensive distributed models (Section 3). In Section 3.3, the parameterisation of these models is detailed.

2 Lumped models

Lateral groundwater flow is not included in the majority of global hydrological models and LSMs. If the goal of modelling is to study groundwater depletion, a volume-based approach is sufficient and lateral flow not essential (Döll et al., 2009, 2012, 2014; Wada et al., 2010; Pokhrel et al, 2012). This section will compare lumped models in the literature by considering how water table dynamics, base flow generation and the groundwater−surface water interaction are simulated. **Table 1** details a selection of lumped models used in either LSMs or global hydrological models.

2.1 WATER TABLE DYNAMICS

The earliest LSMs, as well as most global-scale LSMs, apply a free gravity drainage boundary condition to the bottom of a fixed-depth soil column. This approach assumes that upward flux from the groundwater table is negligible, an assumption that breaks down when the water table is shallow. Including water table dynamics has been shown to improve river discharge simulations (Yeh and Eltahir, 2005; Koirala et al., 2014) and including capillary flux from groundwater increases evapotranspiration, with the global mean simulated to rise by up to 16% (Niu et al., 2007; Anayah et al., 2008; Yeh and Famiglietti, 2009; Koirala et al., 2014).

Gedney and Cox (2003) added an unconfined aquifer layer (12 m thick) under the lowest soil layer of an LSM, but assumed the aquifer to be in equilibrium with the lowest soil layer when the layer was not saturated (i.e. water table depth > 3 m). Their model does not allow for the upward movement of water from the groundwater table, but achieved a better simulation of base flow and wetlands. Niu et al. (2007) developed a simple groundwater model comprising a single unconfined aquifer layer underneath the soil column, which exchanges recharge and capillary flux with the soil column. It explicitly solves the water table depth and then uses it as the lower boundary condition of the model. The model was incorporated into the National Center for Atmospheric Research Community Land Model (Bonan et al., 2002; Niu et al., 2007) and later into the Noah LSM (Niu et al., 2011). Other studies (Liang et al., 2003; Maxwell and Miller, 2005; Yeh and Eltahir, 2005a) incorporated a more realistic representation by explicitly coupling the saturated and unsaturated zones in order to explicitly determine the water table depth. To allow for a deeper water table, they added more nodes or layers to the bottom of the soil column: Yeh and Eltahir (2005a) used 50 soil layers. However, despite the extra layers, the maximum water table depth was still relatively shallow (< 5 m).

More recently, Koirala et al. (2014) incorporated the groundwater representation developed by Yeh and Elathir (2005a) into the MATSIRO (Minimal Advanced Treatments of Surface Integration and Runoff, Takata et al., 2003) LSM. To account for deeper water tables in arid and semi-arid regions, they extended the bottom soil layer to a thickness of 30 m (total model thickness 40 m). That is, the saturated and unsaturated zones become decoupled only when the water table depth is below 40 m. Pokhrel et al. (2015) increased the thickness of the bottom layer of the same model to 90 m to allow for deep water tables resulting from abstraction.

Most lumped groundwater models in LSMs fail to consider groundwater abstraction. Two exceptions are the models developed by Döll et al. (2012, 2014) (WaterGAP, see below for more details) and Pokhrel et al. (2015). Döll et al. (2012, 2014) did not simulate groundwater table dynamics, but abstractions were removed from groundwater storage. Pokhrel et al. (2015) were able to simulate changes in the water table depth caused by pumping. This was undertaken by adding in flows at a $1^{\circ}\times1^{\circ}$ scale. Groundwater withdrawal was estimated as the water demand in excess of surface water availability, with water demand being a combination of consumptive agricultural, domestic and industrial use. Irrigation water demand was calculated with the model's irrigation module and domestic and industrial use were obtained from the AQUASTAT database (http://www.fao.org/nr/water/aquastat/main/index.stm). Pokhrel et al. (2010) evaluated the model's simulated groundwater withdrawal against global-scale groundwater withdrawal data (country based) from Wada et al. (2010), who compiled data from the International Groundwater Resources Assessment Center (https://ggis.un-igrac.org/ggis-viewer/viewer/exploreall/public/default).

2.2 BASE FLOW GENERATION AND PARAMETERISATION

In the WaterGAP global hydrological model used by Döll et al. (2012, 2014), groundwater is represented as a linear reservoir, in which the constant is fixed globally (Müller Schmied et al., 2014). Yeh and Eltahir (2005a) represent groundwater as a non-linear reservoir, having derived a relationship between water table depth and base flow from regression analysis with streamflow as a surrogate for base flow. In a second paper, Yeh and Eltahir (2005b) adapted their model to derive base flow using a statistical-dynamical approach, which they claim accounts for sub-grid heterogeneity in water table depth. Their equation includes the gamma function, which has two parameters, and two other conceptual parameters that cannot be measured and must be calibrated against observed streamflow and inferred base flow information. Yeh and Eltahir (2005b) parameterised the model for locations in Illinois, USA. Koirala et al. (2014) used the same method to study 20 different river basins across the globe. They derived an equation for one parameter based on precipitation and its seasonal variation in Illinois, which was found to be accurate also for uncalibrated basins across the globe. The second parameter was deemed insensitive and fixed (Koirala et al., 2014). In the authors' global model (Pokhrel et al., 2015), the method was simplified to a linear relationship between water table depth and base flow, containing the same two parameters as the previous equation: an outflow constant and a water table depth threshold at which base flow is generated. Pokhrel et al. (2015) used the same values for these parameters as Koirala et al. (2014).

Yeh and Eltahir (2005b) and Koirala et al. (2014) found their models to have low sensitivity to specific yield. Yeh and Eltahir (2005a) fixed specific yield to a value typical of the area they studied, and Niu et al. (2007) and Pokhrel et al. (2015) set it to be globally constant.

Many authors base their derivation of subsurface runoff on the TOPMODEL approach (Beven and Kirkby, 1976), with flow decreasing exponentially as depth to the water table increases (e.g. Gedney and Cox, 2003; Maxwell and Miller, 2005; Niu et al., 2007, 2011). The TOPMODEL-based equation for base flow of Niu et al. (2007, 2011) contains two parameters, which they calibrated globally to runoff data in sensitivity analyses. Maxwell and Miller (2005) had only a single calibration parameter, the saturated hydraulic conductivity at the bottom soil layer.

2.3 GROUNDWATER−SURFACE WATER INTERACTION

One obvious disadvantage of lumped models is their inability to represent groundwater−surface water interactions. All models mentioned in this section have a scheme for generating base flow from the groundwater store, but the model of Döll et al. (2014) is the only one that can simulate recharge from surface water bodies. Their method is, however, very simplified: in areas where precipitation is < 50% of potential evapotranspiration, there is a constant recharge rate per unit area of the surface water body. The rate of recharge varies temporally because the surface water bodies change size with the amount of stored water.

Table 1 Selection of lumped models used in global hydrological models or land surface models

Author	Model	Year	Capillary rise from water table	Water table dynamics	Base flow run off scheme	Groundwater abstraction	Recharge from surface water bodies	Irrigation return flow
Döll et al.	WaterGAP Global Hydrological Model (resolution 0.5˚ x 0.5˚, roughly 55 km x 55 km)	2009			Linear reservoir			×
		2012			Linear reservoir	×		×
		2014			Linear reservoir	×	×	×
Niu et al.	NOAH-MP (part of WRF model)	2007	×	×	TOPMODEL-based, exponential with WTD			
		2011	×	×	TOPMODEL-based, exponential with WTD			
Yeh and Eltahir	Land Surface Transfer Scheme Groundwater (LSXGW)	2005a	×	×	Non-linear reservoir			
Gedney and Cox	Hadley Centre Atmospheric Climate Model (HadAM3) with the Met Office Surface Exchange Scheme (MOSES)	2003		×	TOPMODEL-based, exponential with WTD			
Pokhrel et al.	MATSIRO	2015	×	×	Linear relationship between WTD and base flow	×		×
Maxwell and Miller	Common Land Model (LSM) coupled to ParFLow (groundwater model)	2005	×	×	Simplified TOPMODEL approach, exponential with WTD			
Koirala et al.	MATSIRO	2014	×	×	Statistical-dynamical approach said to account for sub-grid heterogeneity (see Yeh and Eltahir, 2005b)			

WTD, water table depth.

3 Distributed models

Many LSMs ignore lateral groundwater flow on the basis that lateral fluxes between grid cells are very small. Krakauer et al. (2014) showed that significant groundwater flow (> 10% of local recharge or 10 mm/year) occurs over 42% of the global land area at a resolution of 0.1˚ (~1 km), but that this drops to 1.5% at a resolution of 1˚ (~100 km). There are two principal advantages in using a distributed model: (1) a more accurate representation of groundwater−surface water interactions; and (2) a more accurate simulation of water table depth, and thus the effect of groundwater on evapotranspiration and climate (Anyah et al., 2008). There is, however, another reason why groundwater flow is often not represented in LSMs: namely a paucity of hydrogeological data. In this section, distributed models that have been incorporated into LSMs, as well as global distributed groundwater models, are compared. **Table 2** summarises these models.

3.1 MODELS IN LAND SURFACE MODELS

Gutowski et al. (2002) and York et al. (2002) were the first to demonstrate that a distributed groundwater model could be coupled with a single-column land surface−atmosphere model. Gutowski et al. (2002) developed their own simple groundwater model with 1D groundwater flow towards a central river running through the middle of a cell (cell boundaries were no flow boundaries). York et al. (2002) replaced the soil−vegetation and groundwater−surface water modules of the same land surface−atmosphere model with routines integrated into 3D MODFLOW (Harbaugh et al. 2000) for watershed-scale simulations (cell size 50−500 km). The soil−vegetation zone interacts with the aquifer through recharge to the aquifer and evapotranspiration directly from the water table; flow at the catchment outlet is the sum of streamflow and total leakage to or from the aquifer, calculated based on the head difference between the stream and the aquifer.

Fan et al. (2007) incorporated a simple 2D steady state groundwater flow model into an LSM at the continent scale (LEAF-Hydro). The groundwater model assumes that hydraulic conductivity decreases exponentially with depth beneath 1.5 m below ground. The decay factor in the relationship is a function of terrain slope, which the authors parameterised for regolith and bedrock based on concepts of weathering profiles (see Section 2.3). The same authors (Miguez-Macho et al., 2007) extended the model to a transient model with improved treatment of groundwater−surface water interactions. In their original work (Fan et al., 2007), groundwater was discharged to surface water when the water table reached the ground surface and surface water did not discharge to groundwater. This was built upon by Miguez-Macho et al. (2007), who used a statistical approach looking at mean geomorphological parameters across a grid cell (12.5 km resolution), the model resolution being too low for explicit treatment of individual channels. Owing to a lack of geomorphological data, they lumped the river bed hydraulic conductivity, river bed thickness, channel width and channel segment length into one 'river conductance' parameter (as in regional groundwater flow model codes such as MODFLOW), for which an equation was derived comprising equilibrium and dynamic parts. The dynamic part is a function of water table elevation and terrain slope and is calibrated based on river discharge observations. River elevation was fixed from the 'naturally occurring' rivers in the steady state run (1.25 km resolution) (Fan et al., 2007) or at the lowest ground surface elevation, when no river cells occurred within a 12.5 km cell.

Vergnes et al. (2012) added a 2D transient groundwater component to a hydrological model, which they later coupled with an LSM with a scheme for the unsaturated zone (Vergnes et al., 2014). The authors applied the model to France with a parameterisation technique that could be upscaled to the global scale (see Section 2.3). In contrast to most authors, Vergnes et al. (2012, 2014) applied the groundwater model only to the areas where major aquifers had been identified. The representation of the groundwater–surface water interaction is similar to that used in LEAF-Hydro (Miguez-Macho et al., 2007), in that all groundwater cells can exchange water with a 'river' and the rate is determined by a lumped parameter. Vergnes et al. (2012), however, lump only river bed conductivity and thickness into a 'transfer time' parameter and river width and length are calculated with empirical models (see Decharme et al. 2012; Vergnes et al. 2014). The transfer time parameter is dependent on its maximum and minimum values (taken from the literature) and on stream order. Vergnes et al. (2012) note that the system behaviour of the model is more sensitive to the transfer time parameter than to the hydrogeological properties of the aquifer, but also that changes in model performance achieved by varying this parameter are limited compared with the improvement in performance from including groundwater in the hydrological model. In their second paper (Vergnes et al., 2014), which considers capillary flux in the unsaturated zone, the authors reduced capillary flux based on the spatial variability of topography in a grid cell, such that more capillary rise occurs in flatter terrains.

Tian et al. (2012) coupled the AquiferFlow (Wang, 2007) groundwater model to an LSM, using this code to simulate both the saturated and unsaturated zones. They applied the model at the regional scale (grid resolution 3 km, total area ~13 000km^2), incorporating an unconfined aquifer, an aquitard and a confined aquifer. The Heihe river, north-eastern China, and its major tributaries were represented as fixed head boundaries, and the model boundaries as well as the hydraulic conductivity were parameterised through calibration against groundwater level data.

The fully integrated groundwater‒surface water platform ParFlow (Maxwell et al., 2017) has been coupled to the Terrestrial Systems Modeling Platform (TerrSysMP) atmospheric and land surface model (Shrestha et al., 2014), the Weather Research and Forecasting (WRF) atmospheric model (Maxwell et al., 2011), and the regional-scale meteorological model Advanced Regional Prediction System (ARPS) (http://www.caps.ou.edu/ARPS/). A key advantage of ParFlow is its explicit treatment of the groundwater‒surface water interaction, which is either treated as a one-way drainage or parameterised with simple relationships (i.e. a functional relationship between river head and water table depth) in other models. The physically based approach requires little parameterisation, but is computationally extremely expensive, as shown by Maxwell et al. (2015), who applied the model at the continent scale.

3.2 GLOBAL GROUNDWATER MODELS

Fan et al. (2013) presented the first global groundwater model. The model is highly simplified and has a number of limitations. The connection between surface water and groundwater is modelled only implicitly, in that when the groundwater table is above the land surface the excess water is removed. There is no recharge from surface water to groundwater. Moreover, as in their previous work (Fan et al., 2007; Miguez-Macho et al., 2007), no hydrogeological information, such as aquifer thickness or hydraulic conductivity, was used for parameterisation. Instead, a soil database (FAO, 1974) was used to obtain hydraulic conductivity close to the surface and it was then assumed to decrease exponentially. The model is also only steady state, requires calibration to head observations and does not take account of human influences, i.e. pumping, irrigation or drainage.

de Graaf et al. (2015, 2017) presented two global groundwater models of increased sophistication, which built on the group's regional model (Sutanudjaja et al., 2011). They used global datasets on permeability and lithology and a digital terrain model to parameterise hydraulic conductivity and aquifer thickness as well as to delineate confining layers (see Section 2.3) (Gleeson et al., 2011; Hartmann and Moosdorf, 2012). As in LEAF-Hydro (Miguez-Macho et al., 2007), permeability was assumed to decrease exponentially with depth and the rate of decrease is controlled by the terrain slope. The models have a more sophisticated representation of groundwater‒surface water interaction with three variations. (1) For larger rivers (width > 10 m in first paper, >20 m in second), the MODFLOW river (RIV) package was used (same method as used by Miguez-Macho et al., 2007; Vergnes et al., 2012, 2014): that is, the rate of recharge/discharge is based on the head difference between the river and groundwater. The bed resistance parameter (combining bed thickness and conductivity) was set as a constant throughout the model (for which no justification was provided), river width was calculated with an empirical model and river length was assumed equal to the diagonal cell length. (2) Small rivers (width ≤ 10 m in first paper, ≤20 m in second) were simulated by a head dependent leakage function, similar to the MODFLOW drain (DRN) package, with water leaving the groundwater system only when the groundwater level exceeds the surface elevation. (3) An extra term, based on the digital elevation model and estimated storage, was added to account for rivers and springs in mountainous areas for which the model is too coarse to capture.

The first model (de Graaf et al., 2015) was steady state only, comprised a single-layer unconfined aquifer and failed to account for human impacts, i.e. abstraction and irrigation return flow. The model achieved good accuracy (coefficient of determination R^2 = 0.95, regression coefficient α= 0.84) against observed water levels in sedimentary basins, but tended to overestimate groundwater levels in mountainous areas, which the authors attribute to the exclusion of perched aquifers from the model. The second model (de Graaf et al., 2017) was transient, comprised a confined layer as well as the unconfined layer and included abstraction. de Graaf et al. (2017) achieved only a slight improvement in performance by simulating a confined layer. However, they claimed the fact that their estimate of global depletion is closer to that calculated by Konikow (2011) using a volume-based approach demonstrates that the groundwater–surface water interaction is better represented when the confined aquifer is included. However, it should be noted that there is considerable variation between different estimates of global groundwater depletion (from 113 km^3/year [for period 2000–2009] to 330 km^3/year [for year 2000]; Wada et al. 2010; Konikow, 2011; Döll et al., 2014; Pokhrel et al., 2015; de Graaf et al., 2017). Besides the simple approximation of the 3D hydrogeology, necessitated by a lack of data on the global scale, a major limitation of the model is that it is only one-way coupled: that is, the hydrological model is run for the entire simulation period and then time series of surface water levels, net recharge and groundwater abstractions from the hydrological model are passed to the groundwater model. Thus, there is no capillary rise from groundwater and the effects of pumping on surface water levels cannot be included.

Table 2 Distributed groundwater models on a global scale or coupled to land surface models

Author	Year	Groundwater model	Parameterisation		Steady/transient	GW−SW interaction	Resolution	Description
			Hydraulic conductivity and porosity	Aquifer extent and thickness				
Fan et al.	2007	LEAF2-Hydro (LSM)	Vertical K from soil database (FAO, 1974). Anisotropy factor based on soil class to find lateral K. Exponential decrease with depth	Thickness represented by e-folding depth, for which an equation was derived based on slope	Steady	Implicit. When water table is above land surface, water removed as river discharge	1.25 km	North America model
	2013	LEAF2-Hydro	Vertical K from soil database (FAO, 1974). Anisotropy factor based on soil class to find lateral K. Exponential decrease with depth	Thickness represented by e-folding depth, which is function of slope	Steady	Implicit. When water table is above land surface, water removed as river discharge	30 arc-sec (roughly 1 km)	Global model
Miguez-Macho et al.	2007	LEAF2-Hydro (LSM)	Vertical K from soil database (FAO, 1974). Anisotropy factor based on soil class to find lateral K. Exponential decrease with depth	Thickness represented by e-folding depth, which is function of slope	Transient	Statistical approach. River conductance requires calibration against observed river flows	12.5 km	Applied to the USA

Author	Year	Groundwater model	Parameterisation		Steady/transient	GW−SW interaction	Resolution	Description
			Hydraulic conductivity and porosity	Aquifer extent and thickness				
de Graaf	2015	MODFLOW	Gleeson et al. (2011) permeability map	Statistical method for thickness based on topography	Steady	Three different types depending on river size. MODFLOW river (RIV) package used for large rivers	5’	Single, unconfined aquifer. No capillary rise, groundwater pumping or recharge through irrigation return flows. Global model
	2017	MODFLOW	Gleeson et al. (2011) permeability map	Statistical method for thickness based on topography	Transient	Three different types depending on river size. MODFLOW river (RIV) package used for large rivers	5’	Unconfined and confined aquifers. No capillary rise from groundwater. Includes abstraction. Global model
Vergnes	2012	Based on MODCOU (Ledoux et al., 1989)	Transmissivity and effective porosity chosen based on typical values for given lithology	WHYMAP[a], IGME[b] and lithological maps used to delineate main aquifer basins. Slope also used	Transient	All cells are river cells. RC as in MODFLOW. Based on head in river and aquifer. River width calculated with empirical formula.	0.5° and 1/12°	Single layer, unconfined aquifer. GW flow equation solved in spherical coordinates

Author	Year	Groundwater model	Parameterisation		Steady/transient	GW−SW interaction	Resolution	Description
			Hydraulic conductivity and porosity	Aquifer extent and thickness				
	2014	Based on MODCOU (Ledoux et al., 1989)	Transmissivity and effective porosity chosen based on typical values for given lithology	WHYMAP[a], IGME[b] and lithological maps used to delineate main aquifer basins. Slope also used	Transient	All cells are river cells. RC as in MODFLOW. Based on head in river and aquifer. River width calculated with empirical formula.	0.5° and 1/12°	Same as Vergnes et al. (2012), but includes capillary rise and unsaturated zone
Maxwell	2015	ParFlow	Gleeson et al. (2011) permeability maps	Assumed 100 m aquifer thickness everywhere	Steady	Modelled explicitly	1 km (over total area ~6.3 M km^2)	Computationally expensive. No transient dynamics, human activities (e.g. pumping)
York et al.	2002	MODFLOW	From literature (assumed constant across watershed)	Assumed aquifer base	Transient	Proportional to aquifer-river head difference and river conductance (standard MODFLOW)	Variable	Single column LSM-atmosphere model. Watershed scale, single layer, groundwater flow within a cell

Author	**Year**	**Groundwater model**	**Parameterisation**		**Steady/transient**	**GW−SW interaction**	**Resolution**	**Description**
			Hydraulic conductivity and porosity	**Aquifer extent and thickness**				
Gutowski et al.	2002	1D model	From literature (assumed constant)	Assumed from geological knowledge of the area	Transient	River at centre of model cell; river channel extended to aquifer base, no river conductance (flow controlled by aquifer K)	Variable	Single column LSM-atmosphere model. Groundwater flow within cell
Tian et al.	2012	AquiferFlow (unsaturated flow)	Treated as calibration parameters	Borehole logging data	Transient	Fixed head boundaries	3 km	Regional model. Unconfined aquifer, aquitard and confined aquifer. Fully two-way coupled (~ 13,000 km^2)

[a]WHYMAP is available at: http://www.whymap.org.
[b]IGME (International Geological Map of Europe and Adjacent Areas) is available at www.bgr.bund.de.

3.3 PARAMETERISATION

Global datasets that have been used for global groundwater modelling can be found in **Table 3**.

3.3.1 Hydraulic conductivity and porosity

Gleeson et al. (2011) developed global maps of permeability and porosity for consolidated and unconsolidated geological units up to a depth of 100 m (http://spatial.cuahsi.org/gleesont01/). The maps have been used in global-scale (de Graaf et al., 2015, 2017), continent-scale (Maxwell et al., 2015) and regional-scale (Shrestha et al., 2014) models. de Graaf et al. (2015, 2017) used the maps in combination with the high-resolution global lithology map of Hartmann and Moosdorf (2012). Fan et al. (2013), however, mention technical difficulties in using the maps of Gleeson et al. (2011) and instead used the FAO global soil map (FAO, 1974). In their North America model (Fan et al., 2007; Miguez-Macho et al., 2007), the authors used a US soil database that gives conductivity in the vertical direction, so they had to assume an anisotropy factor to determine lateral conductivity and then assumed an exponential decrease in conductivity with depth. Vergnes et al. (2012, 2014) used WHYMAP (http://www.whymap.org), the International Geological Map of Europe (https://www.bgr.bund.de) and a simple lithological map to delineate different geological formations in France and then assumed typical values of hydraulic conductivity and porosity based on the lithology.

Although the assumption of exponentially decreasing hydraulic conductivity with depth – often referred to as the TOPMODEL approach (Beven and Kirkby, 1976) – appears unsatisfactory, it as an assumption that can be found widely in the literature (Niu et al., 2007, 2011; Fan et al., 2007; Miguez-Macho et al., 2007; Fan et al., 2013; Koirala et al., 2014; de Graaf et al., 2015, 2017; Maxwell et al., 2015). Many authors (e.g. de Graaf et al., 2015, 2017; Maxwell et al., 2015) use the e-folding depth[1] (α) first proposed by Fan et al. (2007):

$$K(z) = K_0 e^{-z/\alpha}$$

where K is hydraulic conductivity, K_0 is known hydraulic conductivity at the bottom of the soil layer and z is depth. The authors use the principle that erosion, weathering and deposition determine the decrease in conductivity with depth and that these processes are controlled by slope, climate and bedrock lithology. In order to simplify the approach they base their equation for e-folding on slope alone, such that the steeper the slope, the thinner the regolith. Fan et al. (2007) derived two relationships for e-folding against terrain slope: one for regolith and one for bedrock. The authors mention that the relationships were determined by trial and error, but it is not clear what data were used in deriving them.

3.3.2 Aquifer extent and depth

Shangguan et al. (2017) recently presented the first global map of depth to bedrock (250 m resolution) (https://www.soilgrids.org). Using machine learning algorithms, they derived the map from a global compilation of soil profile data (at 130,000 locations), borehole data (1.6 million locations) and pseudo-observations consisting of remote sensing data, terrain slope and geological maps. The authors warn of low accuracy in extrapolated areas (borehole data are from only eight countries) as well as in areas where depth to bedrock is > 100 m. Cross validation of the data set suggested moderate performance for absolute depth to bedrock.

Pelletier et al. (2016) created a map of average thickness of soil, intact regolith and sedimentary deposits (30 arcsec or ~ 1 km resolution). They used geomorphological models (both process

[1] Here e-folding is used to denote the time interval in which an exponentially growing quantity increases by a factor of e; it is the base-e analog of doubling time.

based and empirical) to calculate these thicknesses based on topographic, climate and geological data. The models were calibrated with data sets from the USA.

de Graaf et al. (2015, 2017) used statistical methods to determine aquifer thickness based on the assumption that mountain ranges have negligible sediment thickness and sediment basins below river valleys contain thicker, more productive aquifers. They used the difference between surface elevation and floodplain elevation within a cell to distinguish between mountain ranges and sediment basins.

Vergnes et al. (2012, 2014) attempted to delineate the main aquifer basins in France using only the WHYMAP global groundwater map, but found it too coarse. They also used the International Geological Map of Europe and a simplified lithological map, removing mountainous areas based on slope.

3.3.3 Categorising confined and unconfined aquifers

de Graaf et al. (2017) delineated confining layers and categorised aquifers into confined and unconfined using information on grain size and sediment properties from the global lithological map GLiM (Hartmann and Moosdorf, 2012). The authors introduced their own method for coastal zones, which are not fully represented in GLiM, classifying coastal zones around large rivers as confined (in total ~11% of global coastline). de Graaf et al. (2017) also made the simplifying assumption that the thickness of the confining layer is always 10% of the estimated aquifer thickness.

Table 3 Global data sets useful in global groundwater modelling

Author	Year	Data	Availability
Gleeson et al.	2011, 2014	Maps of permeability and porosity available at different resolutions	Available at http://crustalpermeability.weebly.com/glhymps.html
Shangguan et al.	2017	Map of depth to bedrock	https://www.soilgrids.org
Pelletier et al.	2016	Maps of average thickness of soil, intact regolith and sedimentary deposits	Available from author on request
FAO	1974	Global soil map	Used by Fan et al. (2007) and Miguez-Macho et al. (2007) to estimate hydraulic conductivity. A digital soil map is available at: https://worldmap.harvard.edu/data/geonode:DSMW_RdY
Bundesanstalt für Geowissenschaften und Rohstoffe		International Geological Map of Europe	Available at: https://www.bgr.bund.de
WHYMAP		Groundwater basins of the world	Available at: https://www.whymap.org/whymap/EN/Home/whymap_node.html
Fan et al.	2013	Map of simulated depth to water table. 1,603,781 groundwater head observations from across the world but predominately in the USA	Online database not found, but data has since been used in other studies (e.g. de Graaf et al., 2015), so can presumably be obtained from the authors
Wada et al.	2010	Global mapping of average recharge and abstraction (in the year 2000, 0.5 x 0.5 degree resolution) with the hydrological model PCR-GLOBWB. Used demand modelling and abstraction data from IGRAC to determine abstraction	
Döll and Fiedler	2008	Global long-term average recharge mapping	
International Groundwater Resources Assessment Centre (IGRAC)	–	Data portal including groundwater abstraction, water level data, etc.	https://ggis.un-igrac.org/ggis-viewer/viewer/exploreall/public/default

4 Recommendations

The method of Döll et al. (2012, 2014) is very attractive in that the simplified lumped approach is appropriate for the amount of data available. However, Döll et al. (2012, 2014) built a global hydrological model to quantify the amount of global groundwater depletion, and it has been shown that simulating the water table depth and capillary rise from the water table has an effect on surface fluxes important in LSMs (Anayah et al., 2008; Kollet and Maxwell, 2008; Koirala et al., 2014; Shrestha et al., 2014). In order to accurately simulate the water table depth, lateral groundwater flow should be included. The method of Maxwell et al. (2015) has the potential to address most limitations in the current groundwater models in LSMs, but the computational expense makes this method unfeasible for the time being. The most practical methods in the literature for including groundwater simulation in an LSM are those of Miguez-Macho et al. (2007) (LEAF-Hydro) and Vergnes et al. (2012, 2014).

Therefore, the recommendation is to test the exponential decay of hydraulic conductivity with depth function incorporated in the LEAF-Hydro approach against existing BGS groundwater flow models, i.e. MaBSWeC (Marlborough and Berkshire Downs and South-west Chilterns; Jackson et al., 2011).

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