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High-resolution hydraulic parameter maps for surface soils in tropical South America

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Abstract

Modern land surface model simulations capture soil profile water movement through the use of soil hydraulics sub-models, but good hydraulic parameterisations are often lacking, especially in the tropics. We present much-improved gridded datasets of hydraulic parameters for surface soil for the critical area of tropical South America, describing soil profile water movement across the region to 30 cm depth. Optimal hydraulic parameter values are given for the Brooks and Corey, Campbell, van Genuchten–Mualem and van Genuchten–Burdine soil hydraulic models, which are widely-used hydraulic sub-models in Land Surface Models. This has been possible through interpolating soil measurements from several sources through the SOTERLAC soil and terrain database and using the most recent pedotransfer functions (PTFs) derived for South American soils. All soil parameter data layers are provided at 15 arcsec resolution and available for download, this being 20 × higher resolution than the best comparable parameter maps available to date. Specific examples are given of the use of PTFs and the importance highlighted of using PTFs that have been locally-parameterised and that are not just based on soil texture. Details are provided specifically on how to assemble the ancillary data files required for grid-based vegetation simulation using the Joint UK Land Environment Simulator (JULES). We discuss current developments in soil hydraulic modelling and how high-resolution parameter maps such as these can improve the simulation of vegetation development and productivity in land surface models.

1 Introduction

Ecosystem water cycles are fundamental to our understanding of how vegetation develops, and how plants respond to periods of high and low water availability. Plants in all ecosystems obtain most of their water through soil, and the study of water movement through the soil matrix has a long history in both ecology and agriculture (Childs, 1969;

Marshall et al., 1996; Leigh, 1999). Soil properties in general are widely recognised as one of the critical factors controlling ecological differences between and within biomes (Phillips et al., 1994, 2004; Leigh, 1999; Aragão et al., 2009; Lloyd et al., 2009; Quesada et al., 2012), as are soil hydraulic properties in particular (Marshall et al., 1996; Tomasella and Hodnett, 1998; Tomasella et al., 2000; Hodnett and Tomasella, 2002).

Soil information is a crucial input for Vegetation and Land Surface Modelling applications: soil properties strongly influence water exchange between the land surface and atmosphere as well as growth processes such as regeneration (e.g. Leigh, 1999; Marthews et al., 2008; Dharssi et al., 2009; Dadson et al., 2011). From a modelling perspective, soil water movement is the result of several overlapping processes, usually simulated by distinct hydrological sub-models (Dadson et al., 2011). At the particle scale within individual soil samples we find that capillary processes dominate (Townend et al., 2001; Hodnett and Tomasella, 2002; Fredlund et al., 2012), at the soil profile or site scale considerations of bypass flow arise (Marshall et al., 1996, 2008), at the landscape scale relative topographic position and groundwater flow are the overriding factors (Dadson et al., 2011) and at the regional scale river routing and other transport processes can overwhelm all other effects (Dadson and Bell, 2010; Dadson et al., 2011). These different processes are all represented within Land Surface Models, although with varying levels of sophistication (e.g. in the JULES model, Best et al., 2011).

For landscape-scale gridded model runs, pre-calculated ancillary files are required to provide the spatially-varying parameter estimates required by hydrological sub-models (e.g. Dharssi et al., 2009; Castanho et al., 2013). The most widely-used, publicly-available set of global ancillary files are currently the IGBP-DIS parameter maps (Global Soil Data Task Group, 2000), however at 5 arcmin resolution (approximately 10 km at the Equator), these maps are now considered fairly coarse (Ke et al., 2012). Land surface models are now being applied at increasingly higher spatial resolution both offline as well as coupled to climate models (Ke et al., 2012), which is necessary to capture the fine-scale dynamics of ecosystem development for realistic modelling of ecosystem productivity and development (Malhi and Wright, 2004; Marthews et al., 2008; Fisher et al., 2008).

Tropical South America is the most intensively-studied tropical region (Malhi and Wright, 2004) so we have taken this region as our focus. The availability of soil-related data for running high-detail simulations has historically been low across the tropics, although most extensive in South America (Tomasella and Hodnett, 1998; Tomasella et al., 2000). In recent years, however, the tropical zone has been recognised as one of the critical "driver" biomes of the world's climate system and the situation is fast improving (Leigh, 1999; Malhi and Wright, 2004; Phillips et al., 2004). Much progress has been made in the availability of high-quality soil information (e.g. the SOil and TERrain database SOTER, Dijkshoorn et al., 2005) and it is now possible to construct ancillary files of much higher resolution and reliability. Additionally, with ever more research groups gaining the capacity to carry out large-scale gridded simulations on a routine basis, there is an increasing need for spatially-explicit parameter maps with which to drive those simulations (e.g. Castanho et al., 2013).

In response to this need, we focus on the profile and sub-profile scale hydraulic models and attempt to improve their parameterisation in the context of tropical South America (Fig. 1). We apply the most up-to-date pedotransfer functions for estimating hydraulic parameters that may be used to produce much more robust simulations of soil water dynamics for this region. Downloadable "model-ready" data grids of parameter values are produced from this analysis, in appropriate formats for use in a wide variety of land surface modelling applications. Finally, we discuss current developments in soil hydraulic modelling in general and identify ways in which these models can benefit from higher-resolution parameter maps such as these.

2 Methods

2.1 Soil hydraulic models

Water movement through saturated soil is described by Darcy's Law, which holds that the vertical rate of water movement is the product of a gradient in hydraulic potential and the soil hydraulic conductivity k (Childs, 1969; Dullien, 1992; Marshall et al., 1996). In unsaturated soil Darcy's Law may be generalised to the Richards Equation (Marshall et al., 1996), which involves k becoming a function of soil matric potential ψ (aka. matric tension, equivalent to capillary pressure head), where ψ is the component of hydraulic potential when any differences in elevation are ignored (Marshall et al., 1996; Mullins, 2001). The relationships between ψ, k and volumetric soil water content θ (unfrozen) can be described by two closely-related curves called the soil water characteristic (SWC) and the soil hydraulic conductivity curve (HCC) (Fig. 2), which together describe the hydraulic model of the soil. In this study we use the four most widely-used soil hydraulic models (Table 1).

2.2 South American soils

We focus on the soils of tropical South America (taken as the area from Panama to the tropic line from Antofagasta, Chile, to São Paulo, Brazil, excluding the Galápagos Archipelago, and the outlying islands Cocos Is., Malpelo Is., Fernando de Noronha, St Peter and St Paul, Trindade and Martim Vaz and Rocas Atoll, Fig. 1). Tropical soils cover a huge range of types and this area is no exception (Ashton, 2004), with parent material varying from the Precambrian rocks of the Guiana and Brazilian Shields to the much younger Cenozoic geology of the Andes and western Amazon (Quesada et al., 2011).

Soil profile measurements were collected from three sources: (i) Quesada et al. (2010) collected data throughout the area as part of the projects RAINFOR http://www.rainfor.org/ and TROBIT www.geog.leeds.ac.uk/groups/trobit/, (ii) Data from the large RADAMBRASIL project (de Negreiros et al., 2009) as supplemented by Cooper et al. (2005) provided good coverage of Brazil and (iii) Data from the International Soil Reference and Information Centre (ISRIC) was used to cover the remaining area (Batjes, 2000). We focussed on surface soil for this analysis because there was insufficient data for adequate mapping of deeper soil: all soil profile data from below 30 cm depth were discarded from these sources, which left a total database of 7620 profile measurements across tropical South America. The soil areas of SOTERLAC (Dijkshoorn et al., 2005) were used, supplemented by data for the same polygons from Batjes (2010).

Over these profile measurements, values for Cation Exchange Capacity (CEC) were used if available from Quesada et al. (2010) and the corresponding polygon in Batjes (2010) (68.8 % of points), otherwise values 10.1 $\mathrm{cmol\,kg^{-1}}$ and 45.7 $\mathrm{cmol\,kg^{-1}}$ were substituted for low and high activity clays, respectively (Hodnett and Tomasella, 2002), following the low- and high-activity categories in Batjes (2010) (Fig. 3). Values for Soil Organic Carbon (SOC) were taken from RADAMBRASIL, Cooper et al. (2005) and Quesada et al. (2010). Values for Dry Bulk Density (DBD) were used if possible from Batjes (2000) (1.2 % of points), otherwise a value of $\max(0.05, 1.578 - (0.054 \cdot (\mathrm{SOC}/10)) - (0.006 \cdot \mathrm{SIPC}) - (0.004 \cdot \mathrm{CLPC}))\,\mathrm{g\,cm^{-3}}$ was substituted (Tomasella and Hodnett, 1998) (a further 90.0 % of points; all variables and units are given in Table 2).

2.3 Spatial analyses

All GIS data analyses were carried out in ArcGIS 10.0 (Esri Inc., Redlands, California). Taking the SOTERLAC polygons as the base areal units, mean values of available soil properties were calculated and assigned to the polygon containing those points. If no profile measurements were available for a particular polygon, a taxotransfer rule was followed with a mean assigned from all profile measurements in the same soil type (taken from Dijkshoorn et al., 2005). In the few cases where no measurements were available from a particular soil type across all tropical South America, a mean

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was assigned from the low- or high-activity soil area of Batjes (2010). This type of calculation is an alternative to smooth interpolation algorithms such as kriging (e.g. Castanho et al., 2013) and is appropriate when extrapolating according to a categorical variable such as soil type which displays spatial step-changes in value.

All soil hydraulic parameters were estimated across tropical South America using pedotransfer functions (PTFs), which are equations used to estimate unavailable soil variables from closely-related and more available soil properties such as texture and dry bulk density (Table 2). PTFs based on tropical soil profiles were used for all quantities (Tomasella and Hodnett, 1998; Hodnett and Tomasella, 2002) except k_{sat}, for which we have been unable to find any continuous, tropically-based PTF (Tomasella and Hodnett, 2004; Rasoulzadeh, 2011; Pan et al., 2012), so we applied instead the most widely-used temperate PTF (from Cosby et al., 1984). Values were calculated on a profile-by-profile basis and then assigned to polygons containing those points using the same rules as for the base measurements. These map layers were then converted to raster format at 15 arcsec resolution and snapped to the HydroSHEDS Digital Elevation Model (DEM) for South America, which is a high-resolution DEM that has had voids and anomalies removed (for details, see Lehner et al., 2006).

3 Results

Parameter maps are available for download for all soil quantities in raster GeoTIFF and NetCDF formats at 15 arcsec resolution (approximately 450 m at the Equator) from http://www.tobymarthews.com/soil-hydraulic-maps.html, with all NetCDF files conforming to Climate and Forecast (CF) conventions. For details of how to use these maps to generate appropriate ancillary data files for grid-based runs using the JULES land surface model, see Appendix A. Note: we present maps of parameter values at regional scale, but these are not parameters for sub-models concerned with regional-scale processes: they are parameters for sub-models concerned with profile-scale processes, intended for use in simulations undertaken across regional-scale domains. Considering the soil quantities from Table 2 in order:

- The saturated hydraulic conductivity k_{sat} is lower across the Amazon Basin than it is across the cerrado (Fig. 4a), with mean across the domain $k_{sat} = 0.008\,mm\,s^{-1}$ ($n = 1902$ SOTERLAC polygons), which is comparable to the $k_{sat} = 0.010\,mm\,s^{-1}$ typical value for a microaggregated clay soil under tropical forest in Panama (Marthews et al., 2008). A reminder: this is a value for surface soil < 30 cm depth (as are all other values given here) and k_{sat} usually decreases with depth (Elsenbeer et al., 1999; Clark and Gedney, 2008).

From the Tomasella and Hodnett (1998) PTFs:

- Parameter b is broadly a measure of the steepness of the SWC (freely-draining sandy soils generally have low values of b (high values of λ) whereas heavy clay soils have high b values). Values derived for tropical South America are uniformly lower (mean = 4.8) than those derived from applying the Cosby et al. (1984) PTF for b (mean = 6.9).

- The air-entry potential ψ_e is the matric potential ψ at which the soil first desaturates when drying after heavy rainfall (i.e. at which the largest pores drain). Values derived for tropical South America are less negative (mean = −1.1 kPa) than those derived from applying the Cosby et al. (1984) PTF for ψ_e (mean = −2.5 kPa).

- The residual soil water content θ_{res} values are generally much lower (mean = $0.04\,cm^3\,cm^{-3}$) than the $0.1866\,cm^3\,cm^{-3}$ value suggested by Tomasella et al. (2000, Table 3) for Brazilian soils.

- The saturated soil water content θ_{sat} values are generally higher (mean = $0.51\,cm^3\,cm^{-3}$) than those derived from applying the corresponding Cosby et al. (1984) PTF (mean = $0.43\,cm^3\,cm^{-3}$) and closely match the $0.502\,cm^3\,cm^{-3}$ value suggested by Tomasella et al. (2000, Table 3) for Brazilian soils.

- Using the parameter correspondences of Morel-Seytoux et al. (1996, Table 1), estimated values for the van Genuchten–Mualem model may also be calculated from these PTFs, giving mean $n_\mathrm{M} = 1.21 \alpha = 0.53\,\mathrm{m}^{-1}$.

From the Hodnett and Tomasella (2002) PTFs:

- The van Genuchten parameter n_M is broadly a measure of how uniform are pore sizes in the soil and this gives n_M values slightly lower (mean = 1.47, Fig. 4b) than the 1.571 value suggested by Tomasella et al. (2000, Table 3) for Brazilian soils. In deeper soil layers, n_M should increase (assuming SOC decreases with depth in the PTF of Hodnett and Tomasella, 2002).
- The van Genuchten parameter α is broadly a measure of how structured the soil is and this gives α values much lower (mean = $3.1\,\mathrm{m}^{-1}$, Fig. 4c) than the $\rho g \cdot (1.0631/1000) = 10.4\,\mathrm{m}^{-1}$ value suggested by Tomasella et al. (2000, Table 3) for Brazilian soils. In deeper soil layers, α should decrease (assuming SOC decreases with depth in the PTF of Hodnett and Tomasella, 2002).
- The residual soil water content θ_res values are much higher (mean = $0.19\,\mathrm{cm}^3\,\mathrm{cm}^{-3}$, Fig. 4d) than those derived from applying the corresponding Tomasella and Hodnett (1998) PTF (mean = $0.04\,\mathrm{cm}^3\,\mathrm{cm}^{-3}$), closely matching the $0.1866\,\mathrm{cm}^3\,\mathrm{cm}^{-3}$ value suggested by Tomasella et al. (2000, Table 3) for Brazilian soils. In deeper soil layers, θ_res should be approximately the same as the surface value (neither DBD nor SOC are in the PTF of Hodnett and Tomasella, 2002).
- The saturated soil water content θ_sat values are slightly lower (mean = $0.48\,\mathrm{cm}^3\,\mathrm{cm}^{-3}$, Fig. 4e) than those derived from applying the corresponding Tomasella and Hodnett (1998) PTF (mean = $0.51\,\mathrm{cm}^3\,\mathrm{cm}^{-3}$) and slightly lower than the $0.502\,\mathrm{cm}^3\,\mathrm{cm}^{-3}$ value suggested by Tomasella et al. (2000, Table 3) for Brazilian soils. In deeper soil layers, θ_sat should decrease (assuming DBD increases with depth in the PTF of Hodnett and Tomasella, 2002).

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Over tropical South America the Hodnett and Tomasella (2002) PTFs give values that appear to be a great improvement over those from Cosby et al. (1984) and Tomasella and Hodnett (1998) (mean SWC curves based on these for standard USDA soil textural classes are given in Hodnett and Tomasella, 2002, if required). Therefore, we recommend these functions whenever data are available for their implementation.

4 Discussion

Our study domain is all of tropical South America, covering all or part of 13 countries. We have also covered the entirety of the Amazon Basin – the most intensely-studied part of the tropics – and the whole of the Amazon forest biome (containing approximately 50 % of global tropical forests and > 50 % of all species that exist on Earth) as well as some 20 % of the world's freshwater resources (Malhi and Wright, 2004). Tropical South America is a critically important region and methods shown to work well here will set the standard for many other areas of the world.

Land Surface Models are being applied at ever increasing spatial resolution as a means to model ecosystem productivity and development as realistically as possible (Malhi and Wright, 2004; Marthews et al., 2008; Fisher et al., 2008; Ke et al., 2012). Characterising soil physical properties is a key element in modelling land surface-atmosphere exchange processes and, therefore, critical to the successful application of coupled Land surface models. The van Genuchten–Mualem model is the current de facto standard soil hydraulic model (Vereecken et al., 2010) and the downloadable data sets provided by this paper are sufficient to parameterise this model across tropical South America, as well as to parameterise three widely-implemented alternatives (Table 1). The maps provided here are of 20x higher resolution than the 5 arcmin resolution IGBP-DIS parameter maps, which are the best previously available (Global Soil Data Task Group, 2000; also see Ashton, 2012) and based on high-quality field data (Quesada et al., 2010) and the most comprehensive soil survey to date for South

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America (SOTERLAC, Dijkshoorn et al., 2005) and the latest pedotransfer functions from field data (Hodnett and Tomasella, 2002).

4.1 Improving model parameterisation

As with all models, soil hydraulics models can only be as good as their parameterisation from field data (see discussions in Vereecken et al., 2010; Zulkafli et al., 2013; Ke et al., 2012). By producing high-quality parameter maps based on pedotransfer functions (PTFs) that are (i) more sophisticated than simple texture-based PTFs (Table 2, specifically including dependence on cation exchange capacity (CEC) which can account at least partially for microaggregation effects) and were (ii) derived from local soil profiles in tropical South America (Hodnett and Tomasella, 2002), these parameter values are as robust as is possible with currently-available data sources. Additionally, the approach and tools we have developed are applicable, with the substitution of locally-derived PTFs, in all other areas of the globe.

Uncertainties around the mean values presented for each hydraulic parameter in Fig. 4 are difficult to estimate because they are calculated from multiple sources, some of which calculated their uncertainties only on a textural class basis (e.g. Hodnett and Tomasella, 2002) and others of which left uncertainty simply unquantified because of the composite nature of their own data sources (e.g. Dijkshoorn et al., 2005; Batjes, 2010). By making the move to spatially-explicit parameter maps, we have quantified between-site variability in these parameter values (which may be calculated for any particular region of interest from the downloadable data layers, with uncertainty ranges for all of tropical South America given in Fig. 4). It has long been known that within-site uncertainty is high for some of these quantities (notably k_{sat}, e.g. see Dirksen, 2001, and θ_{res}, measuring which was described by Durner, 1994 as "a continuous source of vexation"), but we believe that for each quantity presented uncertainty should not exceed 10 % of the regional ranges given in Fig. 4.

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4.2 The wider context: future model development

The soil hydraulic models presented in Table 1 are the current state-of-the-art and all can describe soil water movement well in many soils. However, they are not without limitations and improving these models is an active area of research (e.g. Vereecken et al., 2010; Fredlund et al., 2012). The value of the detailled parameter maps presented in this study would be greatly enhanced if progress could be made on four widely-recognised model improvements in particular:

Improved modelling of macropore flow: Macropore flow is dictated by soil macrostructure rather than microstructure or texture and there is a recognised deficit of knowledge in this area (Lin et al., 2010). For example, the assumption of a unimodal pore-size distribution in standard soil hydraulics models is a significant barrier to further progress (Durner, 1994; Vereecken et al., 2010), especially in areas where microaggregated soils are common (Hodnett and Tomasella, 2002). Modelling macropore flow is especially important for characterising nutrient retention and leaching effects and improving the model representation of real, bimodal soil pore-size distributions could greatly improve the simulation of soil water movement in general (see e.g. Durner, 1994; Kutílek, 2004; Schaap and van Genuchten, 2006; Russell 2010; Schelle et al., 2010; Zeiliguer et al., 2010; Vereecken et al., 2010; Lin et al., 2010). For example, models describing dual-porosity or double-porosity media in the context of flow through fractured rocks may be used to describe bypass and capillary flow in soil columns (Gerke and van Genuchten, 1993; Adler et al., 2005; Guarracino and Monachesi, 2010). In the case where it is unfeasible to move to a multimodal pore-size distribution model, the use of a parameterisation PTF that includes CEC would nevertheless be a significant improvement because that avoids grouping low- and high-activity clays together, resulting in a more realistic placing of the modal pore-size and therefore improved simulations of microaggregated soils where they occur.

Improved modelling of very dry soil: Soil that is exposed to direct sunlight and is within 10 s of cm of the surface can become much drier than permanent wilting point

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(PWP, Fig. 2) and this can occur frequently under natural conditions even in humid forests, for example during short dry spells in a wet season (Marthews et al., 2008). In very dry soil, the liquid water content is no longer continuous so capillary forces become negligible and the strength of adsorptive forces controls water retention (Vereecken et al., 2010; Fredlund et al., 2012). Standard soil hydraulics models do not model these effects well, simply assuming that a certain amount θ_{res} of soil water is "unavailable" to plants (Table 1) even though this approach has clear limitations (Durner, 1994; Vereecken et al., 2010). Improving the model representation of desorptive drying in very dry soil could greatly improve the simulation of soil water movement in general (see e.g. Sillers and Fredlund, 2001; Schelle et al., 2010; Vereecken et al., 2010; Fredlund et al., 2012). This is especially important in environments that are semi-arid or have a pronounced dry season.

Improved modelling of tortuosity: The pore-size distribution (PSD) describes the porosity of the soil (the proportion of micropores and macropores) but how they connect to each other through the soil is described by a pore connectivity model (PCM) and the current almost-universal standard PCM remains that of Mualem (1976). Mualem's PCM included tortuosity as an exponent (parameter L) and work in this area since then has mostly concentrated on deducing or fitting optimal values for L (see Table 1). However, it should be remembered that Mualem (1976)'s study of 45 soils only included 2 clays and 3 clay-loams and no tropical soils at all. Despite its wide use, it seems premature to assume that the Mualem PCM's functional form is optimal for all soil types, and we note that alternative PCMs do exist (see reviews in Dullien, 1992; Kutílek, 2004).

Inclusion of deeper soil layers: The general worldwide lack of soil hydraulic data from below approximately 1.5 m depth, especially regarding hydraulic conductivity, is currently an issue of great importance in soil science and remains a major impediment to the modelling of water movement and uptake in deep-rooted ecosystems such as tropical forests and savannas (Hodnett and Tomasella, 2002). In tropical South America, for example, there remains today an almost complete lack of soil profile data from deep soils (M. Hodnett, personal communication, 2013). Soil profile analysis seldom extends below 1 m as a result of the difficulty in extracting uncompacted and otherwise undisturbed soil cores (see discussion of field methods in Marshall et al., 1996; Dirksen, 2001, although some protocols do extend to deeper layers, e.g. RAINFOR-GEM soil sampling to 4 m, Marthews et al., 2012). In the absence of data from deeper layers, extrapolations are usually made based on topsoil properties (e.g. Clark and Gedney, 2008) despite the much greater compaction of subsoils and the possible presence of markedly different soil layers (e.g. acidic or sodic layers, impermeable layers at shallow depth, Marshall et al., 1996; Lloyd et al., 2009).

In this study we have restricted our analysis to 30 cm soil depth because of the need for a consistent database to extrapolate over a continental area, however understanding the soil sub-surface can nevertheless be crucial. For example, soils in the Acrisols group (Alisols, Luvisols, Lixisols and Acrisols), which cover ~ 30 % of Amazonia, have a typical two layer vertical particle-size distribution, being quite sandy in the top soil and clay rich in the subsoil (usually just below 30 cm). For such soils, the parameter estimates and model fits presented in this study apply only to the upper layers and the effects of the lower layers should be accounted for separately. Despite their sophistication in above-ground aspects, land surface models struggle to simulate systems where sub-surface flow decouples deeper soil layers hydrologically from upper layers, thereby greatly affecting estimates of ecosystem resilience in the long term (see e.g. Elsenbeer et al., 1999 and discussions in Lloyd et al., 2009 and Dadson et al., 2011).

4.3 Spatial patterns of ecosystem function

Soil properties and the dynamics of soil water movement are a fundamental control on the development and productivity of all ecosystems (Lloyd et al., 2009), e.g. Aragão et al. (2009) drew attention to edaphic controls on forest productivity in the Amazon Basin. However, there is a need to separate out which particular soil properties affect vegetation and forest structure in a region such as tropical South America (Quesada et al., 2010, 2011, 2012). Land surface models are an ideal tool for this task because they are process-based and can be used to isolate individual causes and effects, but the current generation of land surface models is not yet sophisticated enough to address this task at continental scales.

For example, it is known that sites in western Amazonia with younger, more fertile soils often have poorer physical properties than sites in central and eastern Amazonia with older, highly weathered soils (e.g. shallower soil depth, lower drainage capacity, Quesada et al., 2010, 2012) and this has been correlated with variation in floristic composition and tree turnover rates across the Amazon Basin (Phillips et al., 2004; Quesada et al., 2012). Similarly, in NE Amazonia generally tree wood is denser and seeds are larger, which has been correlated with the poorer soils that also occur there (ter Steege et al., 2006). However, it is not easy to separate cause and effect here: for example, younger soils are not always more fertile than older soils, and correlations between tree turnover rates and soil physical properties are always potentially confounded with uncontrollable factors such as restricted species ranges and carbon fertilisation rates (see discussions in Phillips et al., 2004 and Ashton, 2004). Some soil properties show a broad east-west variation (e.g. high- vs. low-activity clays, Fig. 3), but other parameters present a more complex spatial pattern – notably soil hydraulic parameters, as shown in this paper (Fig. 4). Spatial patterns such as these imply complex spatial variation in forest dynamics that can only be described very superficially by the current generation of land surface models. Investigations of the causes of these little-studied spatial patterns requires high-resolution ancillary files exactly like those presented here.

5 Conclusions

With recent improvements in the availability of spatial soil data grids (e.g. SOTER, Dijkshoorn et al., 2005) and improved region-specific pedotransfer functions (Tomasella and Hodnett, 2004), it has become possible to generate regional parameter maps for all models of soil hydraulics at high resolution. In this paper we have produced gridded

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datasets for tropical South America that are a step-improvement on the best comparable maps currently publicly-available.

Variation in soil hydraulic parameters may explain less of the known variation in forest dynamics and ecosystem productivity across tropical South America than other factors such as species-specific responses (Leigh, 1999; ter Steege et al., 2006) or climate (e.g. Malhi and Wright, 2004; Lloyd et al., 2009). However, it does seem clear that we currently know far too little about the portion of that variation that is under the control of these parameters. We recognise a pressing need for improved understanding of the various processes that control soil water dynamics, especially soil structural and microstructural aspects of tropical soils. There is also a need for soil datasets that extend profile information to depths of at least 1.5 m in order both to include a greater proportion of the rooting zone and to determine where and when reduced saturated hydraulic conductivity causes lateral flow.

Land Surface Models are experiencing a time of rapid development, but in some ways code development has progressed more quickly than development of the parameterisations on which code simulations are based (e.g. Ke et al., 2012). The time is right for a strong improvement in the quality of parameterisation behind these models, which will lead to much more robust simulations of soil water dynamics and, ultimately, greatly improved vegetation and biome productivity predictions for the tropical zone as a whole.

Appendix A

Parameterising the JULES land surface model

We present here a procedure for assembling the data layers required for parameterising a grid-based vegetation simulation using the JULES land surface model (Best et al., 2011). Using the Brooks and Corey soil hydraulics option as an example (sometimes called "Clapp and Hornberger", see Table 1), several variables must be specified at each point of the simulation grid:

- B is the JULES variable name for the Brooks and Corey parameter $1/\lambda$ (also called b in Campbell, 1974, 1985) and JULES also uses the same variable name to stand for $(1/(n_M-1))$ where n_M is the van Genuchten–Mualem parameter (which makes implicit use of the parameter correspondences given in Table 1). B may be calculated directly from n_M using the HandT layers provided (i.e. parameter maps generated from the Hodnett and Tomasella, 2002 pedotransfer functions (Table 2) available for download from http://www.tobymarthews.com/soil-hydraulic-maps.html).
- SATHH is the JULES variable name for air-entry pressure head h_e and must be in metres. The HandT layers can provide the value of van Genuchten's α in m^{-1} and from that there are two options depending on which of the two parameter correspondences is used (the standard correspondence, which amounts to $h_e = 1/\alpha$, or the correspondence of Morel-Seytoux et al., 1996, which preserves "effective capillary drive" and involves also using the value of B – see Table 1).
- SATCON is the saturated conductivity k_{sat} (in $mm\,s^{-1}$) and is available from the parameter map generated from the Cosby et al. (1984) pedotransfer function (Table 2).
- SM_SAT (= SMVCST or χ_s) is the value $(\theta_{sat} - \theta_{res})$ (Best et al., 2011), which may be calculated directly from two of the HandT layers.
- SM_WILT (= SMVCWT or χ_w) is the value $(\theta_{PWP} - \theta_{res})$, where θ_{PWP} is the water content below which JULES assumes net leaf photosynthesis is zero and may be found by applying the assumed SWC (i.e. following the example calculation given in Table 2 but using the Brooks and Corey equation here instead of van Genuchten).

– SM_CRIT (= SMVCCL or χ_c) is the value ($\theta_c - \theta_{res}$), where θ_c is the water content above which JULES assumes net leaf photosynthesis is at maximum (the "critical point" defined as the volumetric water content at $\psi = -33$ kPa, Best et al., 2011) so this value comes from the assumed SWC in a similar way to the calculation of θ_{PWP}.

Put the same values for all these parameters for all the JULES plant functional types. Global ancillary files are already available for some of these variables, e.g. the UK MET Office Unified Model ancillaries (Jones, 2008), although at lower resolution than the data layers presented in this paper.

Finally, note that in all current versions of JULES the internal soil water content variable for each layer smcl() is actually equal to ($\theta - \theta_{res}$) rather than θ, so if an output variable such as smcTot is selected (= gridbox total soil moisture in column) then θ_{res} must be added to it after the run to return actual soil moisture content values (Jones, 2008; Dharssi et al., 2009; Best et al., 2011). This method becomes increasingly approximate when the residual water content of the soil is high because of uncertainties introduced by other parts of the code where smcl(). is assumed equal to θ (e.g. the calculations of soil thermal conductivity, Best et al., 2011). As has been shown in this study, θ_{res} values as high as 30–40 % are not unusual in many soils (Fig. 4e) and this must be considered in the context of global model runs.

Supplementary material

Soil hydraulic parameter maps for tropical South America, generated from three different pedotransfer functions, are available for download in GeoTIFF and NetCDF formats from http://www.tobymarthews.com/soil-hydraulic-maps.html.

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Table 1. The four most widely-used soil hydraulic models, expressed in terms of ψ (soil matric potential in Pa, n.b. always negative: 0 Pa means saturation and negative values mean increasingly dry soil), θ (the proportion by volume of water in the soil in $cm^3\,cm^{-3}$) and k (soil hydraulic conductivity in $mm\,s^{-1}$, which is the vertical infiltration rate of water into the soil; $k_r = k/k_{sat}$ is the relative hydraulic conductivity). The van Genuchten–Mualem model is currently the de facto standard soil hydraulic model (Vereecken et al., 2010; also see comments in Dirksen, 2001): for reviews of these and other similar models see Shao and Irannejad (1999), Townend et al. (2001), Sillers and Fredlund (2001), Seki (2007), Pertassek et al. (2011) and Fredlund et al. (2012). See Pertassek et al. (2011) and Valiantzas (2011) for the combination Brooks and Corey–Burdine not included here. Matric potential ψ is closely related to (the capillary pressure head of soil water h in m) $= -\psi/(\rho g)$ where ρ = (density of water $1000\,kg\,m^{-3}$) and g = (acceleration of gravity $9.8\,m\,s^{-2}$) (e.g. ψ_e in pressure head units is $h_e = -\psi_e/(\rho g)$ using Pascal's law pressure $= h\rho g$). Other standard quantities: θ_{sat} and k_{sat} are the values of θ and k when the soil is saturated[1] (i.e. all soil pores are water-filled), θ_{res} is the residual soil water content (the soil water content never available to plants, taken here to be the value of θ when the soil is dried to $\psi = -16$ MPa or 1632 m pressure head[2]) and $\Theta = \frac{\theta - \theta_{res}}{\theta_{sat} - \theta_{res}}$ is the normalised volumetric water content (aka. effective saturation S_e).

Model	Soil water characteristic (SWC)	Hydraulic conductivity curve (HCC)[3] ($k_r = k/k_{sat}$)	Parameters	Notes
Brooks and Corey – Mualem (BandC-M)[4,5]	$\Theta = \begin{cases} 1 & \text{for } \psi \geq \psi_e \\ \left(\frac{\psi}{\psi_e}\right)^{-\lambda} & \text{for } \psi < \psi_e \end{cases}$	$k_r = \begin{cases} 1 & \text{for } \psi \geq \psi_e \\ \Theta^{L+2+\frac{2}{\lambda}} = \left(\frac{\psi}{\psi_e}\right)^{-\lambda(L+2)-2} & \text{for } \psi < \psi_e \end{cases}$	Pore-size distribution index λ (≥ 0, no units) Air-entry/saturation potential or bubbling pressure ψ_e (< 0, Pa) Pore tortuosity parameter L (no units)	Brooks and Corey (1964) presented the $L = 1$ case and this is implemented in JULES (Best et al., 2011). Mualem (1976) generalised this model to $L \neq 1$ and suggested $L = 0.5$[6].
Campbell (– Mualem)[4]	$\Theta = \frac{\theta}{\theta_{sat}} = \begin{cases} 1 & \text{for } \psi \geq \psi_e \\ \left(\frac{\psi}{\psi_e}\right)^{-\frac{1}{b}} & \text{for } \psi < \psi_e \end{cases}$	$k_r = \begin{cases} 1 & \text{for } \psi \geq \psi_e \\ \Theta^{2b+3} = \left(\frac{\theta}{\theta_{sat}}\right)^{2b+3} = \left(\frac{\psi}{\psi_e}\right)^{-2-\frac{3}{b}} & \text{for } \psi < \psi_e \end{cases}$	(Inverted) pore-size distribution index b (≥ 0, no units) Air-entry/saturation potential or bubbling pressure ψ_e (< 0, Pa)	Derived from BandC-M by taking $\theta_{res} = 0\,\mathrm{cm^3\,cm^{-3}}$, $b = 1/\lambda$ and $L = 1$[6] (Campbell, 1974, 1985).
van Genuchten – Mualem (vanG-M)[4,7]	$\Theta = (1+A)^{-m_M}$	$k_r = \Theta^L \left(1 - \left(1 - \Theta^{\frac{1}{m_M}}\right)^{m_M}\right)^2 = (1+A)^{-Lm_M}\left(1 - \left(\frac{A}{1+A}\right)^{m_M}\right)^2$	Pore-size distribution index n_M (≥ 1, no units)[8] Inverse value of the bubble point potential α (> 0, $\mathrm{m^{-1}}$) Pore tortuosity parameter L (no units) n.b. $m_M = 1-(1/n_M)$ so $n_M = 1/(1-m_M)$	Van Genuchten (1980) presented the $L = 0.5$ case ($A = (\alpha h)^{n_M}$, $h = \frac{-\psi}{\rho g}$), but with Mualem (1976)'s generalisation we can have $L \neq 0.5$ as well[6]. This model is implemented in ORCHIDEE (de Rosnay et al., 2000) and also in JULES (Jones, 2008; Dharssi et al., 2009; Best et al., 2011).
van Genuchten – Burdine (vanG-B)[7]	$\Theta = (1+A)^{-m_B}$	$k_r = \Theta^2 \left(1 - \left(1 - \Theta^{\frac{1}{m_B}}\right)^{m_B}\right) = (1+A)^{-2m_B}\left(1 - \left(\frac{A}{1+A}\right)^{m_B}\right)$	Pore-size distribution index n_B (≥ 2, no units) α (> 0, $\mathrm{m^{-1}}$) n.b. $m_B = 1-(2/n_B)$	(where $A = (\alpha h)^{n_B}$, $h = \frac{-\psi}{\rho g}$) from Burdine (1953) and van Genuchten (1980)

[1] In any particular soil layer, θ_{sat} is usually 5–10 % smaller than the total porosity because of entrapped or dissolved air (van Genuchten et al., 1991).
[2] This residual matric potential $\psi = -16\,\mathrm{MPa}$ is approximately the limit of field conditions for desert soils (cf. discussion in Durner, 1994 on how to measure θ_{res}).
[3] k may be restricted to $\geq 10^{-9}\,\mathrm{mm\,s^{-1}}$ on physical grounds (e.g. Marthews et al., 2008).
[4] Correspondence between BandC-M and vanG-M: if you have values for the BandC-M parameters λ and ψ_e (assuming $L = 1$) then Morel-Seytoux et al. (1996) proposed the following to calculate equivalent vanG-M parameters n_M and α: $m_M = \frac{\lambda}{\lambda+1}$ and $\alpha = \frac{\rho g}{-\psi_e} \cdot \frac{147.8+8.1p+0.092p^2}{\frac{2p(p-1)}{p+3}(55.6+7.4p+p^2)}$ (where $p = 3 + \frac{2}{\lambda}$). Alternatively, Dharssi et al. (2009) and Valiantzas (2011) gave a simpler equivalence scheme: $m_M = \frac{\lambda}{\lambda+1}$ and $\alpha = \frac{\rho g}{-\psi_e}$ (equivalently $\lambda = n_M - 1$ or $b = \frac{1}{n_M-1}$ and $h_e = \frac{1}{\alpha}$), but this correspondence should be considered approximate because it does not preserve the "effective capillary drive" implied by the soil model (Morel-Seytoux et al., 1996). Because Campbell's model is a special case of BandC-M, these correspondence schemes also apply between Campbell and vanG-M.
[5] Brooks and Corey's model is called the "Clapp and Hornberger model" at the European Centre for Medium-Range Weather Forecasts (ECMWF 2009) and by many users of the JULES land surface model (e.g. Jones, 2008; Dharssi et al., 2009; Best et al., 2011) because of the contribution Clapp and Hornberger (1978) made in popularising it.
[6] $L = 0.5$ is currently a hard-wired default in the Met Office operational global Unified Model. Other values are also in use such as the ECMWF (2009, Table 7.6) values in the range −2.342 to 2.500, although values of $L < 0$ may not be physically realistic because they imply that the connectivity of water-filled pores increases with decreasing soil moisture content (see discussions in Wösten et al., 1999; Schaap and Leij, 2000; Vereecken et al., 2010). Slightly inconsistently, Brooks and Corey soil hydraulics are also an option in the Unified Model which implicitly assumes $L = 1$ (see above). Finally, note that this tortuosity parameter L is not directly related to the Kozeny tortuosity parameter k of the Kozeny–Carman equation (Dullien, 1992).
[7] Correspondence between vanG-M and vanG-B: Use $n_B = n_M + 1$ and $m_B = \frac{m_M}{2-m_M}$ and the α parameter is identical between these two models (van Genuchten, 1980).
[8] Soil microporosity and macroporosity increase as $(m_M n_M)$ and (m_M/n_M) increase, respectively (Durner, 1994), so the proportion of micropores should be highest for high values of n_M and the proportion of macropores should be maximal for n_M close to 2.

Table 2. Three sets of pedotransfer functions (PTFs) for deriving values for the parameters of soil hydraulic models from more easily-measured soil properties, of which the Hodnett and Tomasella (2002) PTFs are the ones we recommend for general use in tropical South America (for reviews of many more PTFs see Tomasella and Hodnett, 1998, 2004; Wösten et al., 1999; Tomasella et al., 2000, 2003; Vereecken et al., 2010; Rasoulzadeh, 2011; Pan et al., 2012). Independent variables are CLPC = clay fraction in %, SAPC = sand fraction in %, SIPC = silt fraction in % = 100 – CLPC – SAPC, DBD = (dry bulk density in $\mathrm{Mg\,m^{-3}}$ or, equivalently, $\mathrm{g\,cm^{-3}}$), CEC = (cation exchange capacity in $\mathrm{cmol\,kg^{-1}}$), SOC = (soil organic carbon content in $\mathrm{g\,C\,kg^{-1}}$ soil), pH = (hydrogen ion activity, dimensionless). For Brooks and Corey and van Genuchten, there is some evidence that L varies between soil categories (e.g. Wösten et al., 1999; Schaap and Leij, 2000; ECMWF, 2009, Table 7.6), but we are not aware of any continuous, tropically-based PTF for this parameter and suggest simply to assume $L = 0.5$. It should be remembered that the relationship between soil pore-size and particle-size distributions is only approximate, especially in highly-leached soils and low-activity clays: SWC curves are closely related to the pore-size distribution, whereas soil texture is related to particle sizes, so there is inherent uncertainty involved in using PTFs based only on texture (Vereecken et al., 2010). Note also that these are all continuous PTFs which are considered to be more robust than class PTFs where only the soil category (e.g. silty clay loam, sandy clay) is required to obtain estimates of the desired soil variables (Wösten et al., 1999; Nemes et al., 2001; Hodnett and Tomasella, 2002), not least because class PTFs are dependent on the classification used which may not be optimal especially for tropical soils (Twarakavi et al., 2010). Finally, remember also that all pedotransfer functions are approximations: direct measurement of these parameters in situ is always preferable (curves fitted to soil moisture measurements, e.g. Marthews et al., 2008, Fisher et al., 2008), but unfortunately often difficult or unfeasible (Tomasella and Hodnett, 1998; Townend et al., 2001; Hodnett and Tomasella, 2002; Zulkafli et al., 2012).

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Source	PTFs (see Table 1 for quantity definitions)
Cosby et al. (1984)[2] Non-tropical and texture-based, but widely-used	(λ, dimensionless) = $1/(3.10 + 0.157 \cdot \text{CLPC} - 0.003 \cdot \text{SAPC})$ (b, dimensionless) = $1/\lambda = 3.10 + 0.157 \cdot \text{CLPC} - 0.003 \cdot \text{SAPC}$ (ψ_e in[3] Pa) = $-0.01 \cdot (10\hat{}(2.17 - (0.0063 \cdot \text{CLPC}) - (0.0158 \cdot \text{SAPC}))) \cdot (1000 \cdot 9.80665)$ (ψ_e in m of pressure head, also called h_e) = $-$ (ψ_e in Pa)/(ρg) = $0.01 \cdot (10\hat{}(2.17 - (0.0063 \cdot \text{CLPC}) - (0.0158 \cdot \text{SAPC})))$ (θ_{sat} in[1] $\text{cm}^3\,\text{cm}^{-3}$) = $0.01 \cdot (50.5 - 0.037 \cdot \text{CLPC} - 0.142 \cdot \text{SAPC})$ (k_{sat} in[4] $\text{mm}\,\text{s}^{-1}$ or, equivalently, $\text{kg}\,\text{m}^{-2}\,\text{s}^{-1}$) = $(25.4/(60 \cdot 60)) \cdot (10\hat{}(-0.60 - (0.0064 \cdot \text{CLPC}) + (0.0126 \cdot \text{SAPC})))$
Tomasella and Hodnett (1998) Tropical, but still texture-based	(λ, dimensionless) = $\exp(-(1.197 + (0.00417 \cdot \text{SIPC}) - (0.0045 \cdot \text{CLPC}) + (0.000894 \cdot \text{SIPC} \cdot \text{CLPC}) - (0.00001 \cdot \text{SIPC} \cdot \text{SIPC} \cdot \text{CLPC})))$ (b, dimensionless) = $1/\lambda = \exp(1.197 + (0.00417 \cdot \text{SIPC}) - (0.0045 \cdot \text{CLPC}) + (0.000894 \cdot \text{SIPC} \cdot \text{CLPC}) - (0.00001 \cdot \text{SIPC} \cdot \text{SIPC} \cdot \text{CLPC}))$ (ψ_e in Pa) = $-1000 \cdot (0.285 + (0.000733 \cdot \text{SIPC} \cdot \text{SIPC}) - (0.00013 \cdot \text{SIPC} \cdot \text{CLPC}) + (0.0000036 \cdot \text{SIPC} \cdot \text{SIPC} \cdot \text{CLPC}))$ (ψ_e in m of pressure head, also called h_e) = $-$(ψ_e in Pa)/(ρg) = $(1/9.80665) \cdot (0.285 + (0.000733 \cdot \text{SIPC} \cdot \text{SIPC}) - (0.00013 \cdot \text{SIPC} \cdot \text{CLPC}) + (0.0000036 \cdot \text{SIPC} \cdot \text{SIPC} \cdot \text{CLPC}))$ (θ_{sat} in[1] $\text{cm}^3\,\text{cm}^{-3}$) = $0.01 \cdot (40.61 + (0.165 \cdot \text{SIPC}) + (0.162 \cdot \text{CLPC}) + (0.00137 \cdot \text{SIPC} \cdot \text{SIPC}) + (0.000018 \cdot \text{SIPC} \cdot \text{SIPC} \cdot \text{CLPC}))$ (θ_{res} in $\text{cm}^3\,\text{cm}^{-3}$) = $0.01 \cdot \max(0, -2.094 + (0.047 \cdot \text{SIPC}) + (0.431 \cdot \text{CLPC}) - (0.00827 \cdot \text{SIPC} \cdot \text{CLPC}))$
Hodnett and Tomasella (2002, Table 8), Tomasella and Hodnett (2004) Tropical and more sophisticated than texture-based PTFs	(n_M, dimensionless) = $\exp((62.986 - (0.833 \cdot \text{CLPC}) - (0.529 \cdot (\text{SOC}/10)) + (0.593 \cdot \text{pH}) + (0.007 \cdot \text{CLPC} \cdot \text{CLPC}) - (0.014 \cdot \text{SAPC} \cdot \text{SIPC}))/100)$ (α in m^{-1}) = $(1000 \cdot 9.80665)/(1000/\exp((-2.294 - (3.526 \cdot \text{SIPC}) + (2.440 \cdot (\text{SOC}/10)) - (0.076 \cdot \text{CEC}) - (11.331 \cdot \text{pH}) + (0.019 \cdot \text{SIPC} \cdot \text{SIPC}))/100))$ (α in Pa^{-1}) = (α in m^{-1})/(ρg) = $0.001 \cdot \exp((-2.294 - (3.526 \cdot \text{SIPC}) + (2.440 \cdot (\text{SOC}/10)) - (0.076 \cdot \text{CEC}) - (11.331 \cdot \text{pH}) + (0.019 \cdot \text{SIPC} \cdot \text{SIPC}))/100)$ (θ_{sat} in[1] $\text{cm}^3\,\text{cm}^{-3}$) = $0.01 \cdot (81.799 + (0.099 \cdot \text{CLPC}) - (31.42 \cdot \text{DBD}) + (0.018 \cdot \text{CEC}) + (0.451 \cdot \text{pH}) - (0.0005 \cdot \text{SAPC} \cdot \text{CLPC}))$ (θ_{res} in $\text{cm}^3\,\text{cm}^{-3}$) = $0.01 \cdot (22.733 - (0.164 \cdot \text{SAPC}) + (0.235 \cdot \text{CEC}) - (0.831 \cdot \text{pH}) + (0.0018 \cdot \text{CLPC} \cdot \text{CLPC}) + (0.0026 \cdot \text{SAPC} \cdot \text{CLPC}))$

Example calculation using values from Tambopata forest plot TAM-05, Peru (12°49′49″ S, 69°16′14″ W, Phillips et al., 2004). At this location, the PTFs from Hodnett and Tomasella (2002) give the following values: $n_M = 1.08$, $\alpha = 7.62\,\text{m}^{-1}$, k_{sat} (from Cosby et al., 1984) = $0.0024\,\text{mm}\,\text{s}^{-1}$, $\theta_{sat} = 0.86\,\text{cm}^3\,\text{cm}^{-3}$ and $\theta_{res} = 0.25\,\text{cm}^3\,\text{cm}^{-3}$ (high values because the soil is a clay and the six soil profile measurements for Tambopata in our database all had high SOC (mean = $501\,\text{g}\,\text{kg}^{-1}$) and low CEC (mean = $4.03\,\text{cmol}\,\text{kg}^{-1}$), indicating likely microaggregation).

Put in $\psi = -1\,500\,000$ Pa for permanent wilting point (see Table 1 for equations, taking $L = 0.5$): $h = -\psi/(1000 \cdot 9.80665) = 152.9$ m, $A = (\alpha h)^n = 2117.3$, $\Theta = (1+A)\hat{}((1/n)-1) = 0.55$ so $\theta_{PWP} = (\Theta \cdot (\theta_{sat} - \theta_{res})) + \theta_{res} = 0.58\,\text{cm}^3\,\text{cm}^{-3}$.

Put in $\psi = -10\,000$ Pa for field capacity (we use the standard -10 kPa not -33 kPa, following Marshall et al., 1996; Townend et al., 2001; Tomasella and Hodnett, 2004): $h = 1.02$ m, $A = 9.2$, $\Theta = 0.83$ so $\theta_{FC} = 0.76\,\text{cm}^3\,\text{cm}^{-3}$.

So the Available Water Capacity of this soil down to 30 cm depth is AWC = $(\theta_{FC} - \theta_{PWP}) \cdot 0.30 = (0.76 - 0.58) \cdot 0.30 = 0.054$ m water or 54 mm water. The Profile Available Water Capacity of this soil down to the 1.7 m depth of the soil column (Townend et al., 2001), assuming that this soil is shallow enough for these surface parameters to be assumed constant over the soil column, is PAWC = $(\theta_{FC} - \theta_{PWP}) \cdot \text{depth} = (0.76 - 0.58) \cdot 1.7 = 310$ mm water or 180 mm water m^{-1} soil over the soil column (slightly lower than might be expected given the high θ_{sat} value).

Values for k_r and k may be calculated similarly and the Morel-Seytoux et al. (1996) parameter correspondence (Table 1) used to generate corresponding parameters for the Brooks and Corey model if necessary. All values should ideally be checked for consistency with the global mean values of UNSODA (Nemes et al., 2001).

[1] Using the standard relationship $\theta_{sat} = 1 - (\text{DBD}/\rho_p)$ with θ_{sat} in $\text{cm}^3\,\text{cm}^{-3}$, DBD in $\text{g}\,\text{cm}^{-3}$ and soil particle density ρ_p in $\text{g}\,\text{cm}^{-3}$ (Marshall et al., 1996) and an assumed approximation for ρ_p (e.g. $\rho_p = ((1.25 \cdot \text{SOM}) + (2.65 \cdot (100 - \text{SOM})))/100\,\text{g}\,\text{cm}^{-3}$ where SOM = (SOC/10)/0.58 %), it is possible to replace the given PTFs for θ_{sat}, however uncertainty in the value of ρ_p is generally high at continental scales so this approach was not used here.

[2] In the JULES land surface model these equations were previously misquoted with exponentials exp() rather than powers of ten 10ˆ() (see Jones, 2008; Dharssi et al., 2009).

[3] For ψ_e many sources use units $\text{J}\,\text{kg}^{-1}$ or kPa which are equivalent to each other (see Marshall et al., 1996, p. 37) but a factor of 1000 different from these units.

[4] For k_{sat} many sources use units $\text{cm}\,\text{s}^{-1}$ or $\text{kg}\,\text{s}\,\text{m}^{-3}$ which are equivalent to each other (see Marshall et al., 1996, p. 37) but a factor of 10 different from these units.

Fig. 1. Physical map of tropical South America (based on the HydroSHEDS Digital Elevation Model, Lehner et al., 2006). National borders and the outline of Amazonia sensu stricto are shown (Eva et al., 2005).

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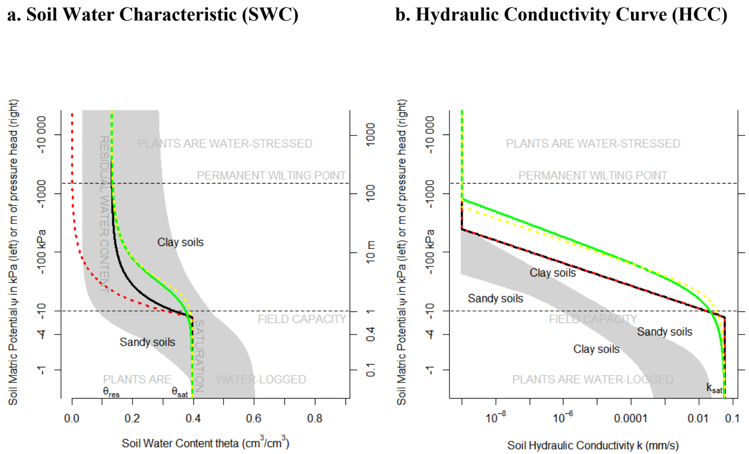

Fig. 2. An example soil water characteristic (SWC) **(a)** and corresponding hydraulic conductivity curve (HCC) **(b)** (Childs, 1969; Townend et al., 2001; Dirksen 2001) for a silt loam soil with no shrinkage or hysteresis (example from van Genuchten, 1980). When the soil is saturated with water (e.g. just after heavy rain), $\theta = \theta_{sat}$ and $k = k_{sat}$ (see Table 1 for these quantities). As the soil begins to dry and becomes unsaturated, the soil moisture content and conductivity fall rapidly until gravity drainage ceases and field capacity is reached (usually a few days after rain) (Fredlund et al., 2012). If drying continues, plants become increasingly water-stressed and will begin to exhibit damage at permanent wilting point ($\psi = -1500$ kPa or 153 m pressure head) where $\theta = \theta_{PWP}$. The solid black, broken red, solid green and broken yellow curves are the appropriate Brooks and Corey, Campbell, van Genuchten–Mualem and van Genuchten–Burdine models for this soil, respectively (Table 1; Brooks and Corey coincides exactly with Campbell on **b**). The grey bands show the range of values across tropical soil categories considered by Hodnett and Tomasella (2002), assuming k_{sat} values from the pedotransfer function of Cosby et al. (1984) (Table 2), from which it may be seen that this example soil has a relatively low saturated water content but high hydraulic conductivity. Unrealistically small values of k_{unsat} are avoided by restricting k_{unsat} to $\geq 10^{-9}$ mm s^{-1} (Marthews et al., 2008).

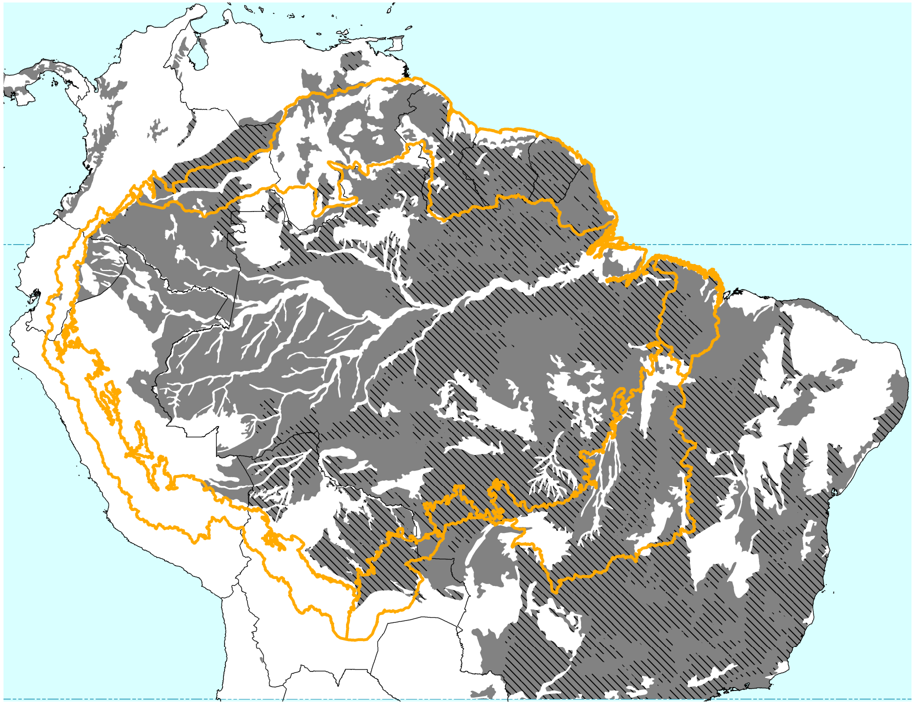

Fig. 3. Map of low-activity soils in tropical South America (57.0 % of the area, Batjes, 2010) (▇), overlaid by shading indicating the occurrence of Ferralsols (= Oxisols, 29.2 %, Dijkshoorn et al., 2005) (╲╲╲). National borders and the outline of Amazonia sensu stricto are also shown (Eva et al., 2005).

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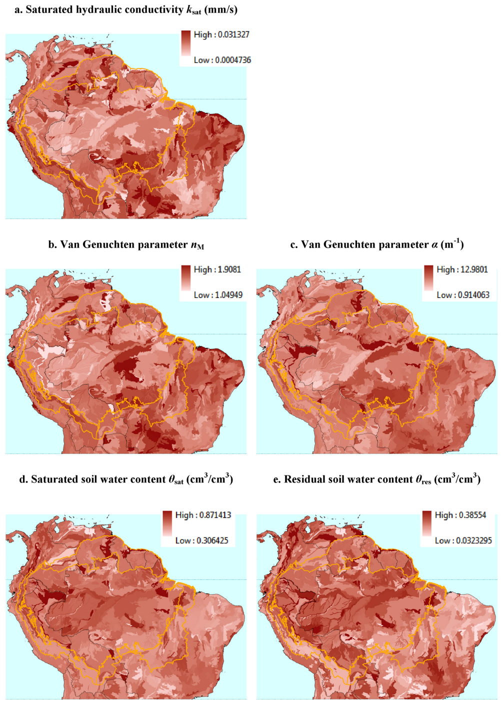

Fig. 4. Soil hydraulic parameter maps for tropical South America derived from the pedotransfer functions of Cosby et al. (1984) **(a)** and Hodnett and Tomasella (2002) **(b–e)**: **(a)** saturated hydraulic conductivity k_{sat} (mm s^{-1}), **(b)** Van Genuchten parameter n_M, **(c)** Van Genuchten parameter α (m^{-1}), **(d)** saturated soil water content θ_{sat} (cm^3 cm^{-3}) and **(e)** residual soil water content θ_{res} (cm^3 cm^{-3}). National borders and the outline of Amazonia sensu stricto are also shown (Eva et al., 2005). For example, the calculated values given for Tambopata (Table 2) were from an area in SE Peru with unusually high residual and saturated soil water content in its soils and maps **(c)** and **(d)** give an indication of where soils of similar hydraulic properties may be found elsewhere in the region.

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