52 Changes in soil physical properties after conversion of tropical montane cloud forest to pasture in northern Costa Rica

C. Tobón
Universidad Nacional de Colombia, Medellín, Colombia

L. A. Bruijnzeel and K. F. A. Frumau
VU University, Amsterdam, The Netherlands

J. C. Calvo-Alvarado
Instituto Tecnologíco de Costa Rica, Cartago, Costa Rica

ABSTRACT

Within the framework of a larger project studying the hydrologic impacts of converting tropical montane cloud forest to pasture in the Tilarán range of northern Costa Rica, physical and hydraulic properties of various volcanic soils were compared in two small watersheds covered with mature lower montane cloud forest and pasture, respectively. *In situ* and laboratory experiments were conducted to determine trends in soil texture, bulk density, porosity, water retention characteristics, infiltration, and (un)saturated hydraulic conductivities with depth under the two types of land cover. Despite their predominantly sandy texture, the soils were rich in organic matter and non-crystalline material such as allophane. Bulk densities were very low and similar between sites for corresponding soil horizons, except for the pasture top-soil which was more compacted, particularly on cow trails. Soil porosity was very high throughout the profile and dominated by macro- and mesopores, again with the exception of the pasture top-soil and the cow trails. Water retention at a suction of 1500 kPa (permanent wilting point) was very high, except in gravelly C-horizons which had low retention capacity. Amounts of plant-available water (i.e. held at suctions between 10 and 1500 kPa) were also high. Surface infiltration rates were relatively high and dominated by "bypass" flow via macropores in the gravelly horizon (at 20–30 cm depth). Spatial variability in infiltration rates was high in the pasture but less in the forest. Saturated hydraulic conductivity at the soil surface was high in general but considerably reduced in the pasture. Values obtained by well permeametry (Guelph permeameter) were higher and considered more representative than those obtained by tension infiltrometry. Unsaturated hydraulic conductivities were also quite high, even at high soil water suctions, except for the gravelly C-horizon, where conductivity decreased almost linearly with decreasing water content. Overall, the results indicate that converting cloud forest to grazed pasture diminishes infiltration of water into the soil due to increased top-soil density and reduced porosity, potentially resulting in increased overland flow and runoff response to rainfall.

INTRODUCTION

Montane cloud forests are important for their high water yield due to a combination of reduced evaporation losses and extra inputs afforded by the interception of wind-driven rain and fog (Zadroga, 1981; McJannet *et al*., this volume), for erosion control on steep slopes, and for their unique biodiversity (Hamilton *et al*., 1995; Kappelle and Brown, 2001) As populations depending on the clean water emanating from these

Tropical Montane Cloud Forests: Science for Conservation and Management, eds. L. A. Bruijnzeel, F. N. Scatena, and L.S. Hamilton. Published by Cambridge University Press.

forests increase, so does the importance of cloud forest conservation and understanding the processes underlying their hydrologic behavior. Whilst progress has been made in recent years with the quantification of horizontal precipitation inputs to cloud forests (Juvik and Nullet, 1995; Holwerda *et al*., 2006; McJannet *et al*., 2007b; Giambelluca *et al*., this volume; Schmid *et al*., this volume), and evaporative losses from cloud forests (Santiago *et al*., 2000; Holwerda, 2005; Motzer *et al*., 2005; McJannet *et al*., 2007a; Giambelluca *et al*., 2009), their soil water dynamics have remained rather poorly documented (Herrmann, 1971; Hafkenscheid *et al*., 2002; García-Santos, 2007; Bogner *et al*., 2008; Schrumpf *et al*., this volume).

Soils under mature tropical forests are generally well aggregated and typically have high infiltration capacity and low (if any) infiltration-excess overland flow (Bonell, 2005) Upon conversion to pasture or annual cropping, considerable changes in soil properties usually occur, particularly after mechanized logging and clearing (Lal, 1987; Schwartz *et al*., 2000; Grip *et al*., 2005; Huwe *et al*., 2008; Zimmermann and Elsenbeer, 2008) Cloud forest clearing on steep slopes often does not involve heavy machinery, whereas rainfall intensities tend to be lower at higher elevations, whereas, in addition, bracken ferns often invade freshly burned montane areas, thereby affording some surface protection against erosion (Aide *et al*., this volume). As such, the surface impacts of cloud forest clearing may not be all that pronounced (e.g. Edwards, 1979; Duisberg-Waldenberg, 1980), although compaction of pasture top-soils by roaming cattle can be substantial, as demonstrated by the work of Zimmermann and Elsenbeer (2008) on soils derived from metamorphic rocks in southern Ecuador. Volcanic soils represent a special case in that these may show irreversible physical changes upon drying (Kubota, 1976) However, there is comparatively little information on the physical and hydraulic properties of volcanic soils under pristine cloud forest conditions and changes therein after forest conversion (Regalado and Ritter, 2006; García-Santos, 2007; Podwojewski *et al*., 2008; Schrumpf *et al*., this volume).

Applications of watershed hydrologic response models have increased steadily and these require spatially representative parameters describing the physical and hydraulic properties of the soils (Saghafian *et al*., 1995; Woolhiser *et al*., 1996; Parasuraman *et al*., 2006; Lazarovitch *et al*., 2007; cf. Huwe *et al*., 2008) To address this lack of information regarding the soil hydrologic impacts of tropical montane cloud forest conversion to pasture, detailed field and laboratory measurements were made of the texture, bulk density, porosity, water retention, infiltration, and hydraulic conductivity of volcanic soils in two small watersheds with mature cloud forest and pasture in the Tilarán range of northern Costa Rica.

STUDY AREA

This study was carried out in two small first-order watersheds in the San Gerardo area on the wet Atlantic slopes of the Tilarán range near Santa Elena, northern Costa Rica (10° 21′ 33″ N, 84° 48′ 05″ W). The cloud forest watershed (3.5 ha) was situated between 1450 and 1600 m.a.s.l. and the pasture watershed (8.7 ha) between 1520 and 1620 m.a.s.l. The area is characterized by high rainfall (4400–6000 mm year^{-1}) and strong winds, and receives considerable amounts of horizontal precipitation, bringing the total annual input to ~9000 mm (Frumau *et al*., 2006) Whilst the drier Pacific slopes experience a well-defined dry season from January until May, the dry season is much less pronounced on the wet Atlantic slopes (Zadroga, 1981; Clark *et al*., 2000) Average temperature in 2003 was 17.0°C (monthly range 15.4–17.7°C). Relative humidity was generally above 90% and foggy conditions prevailed for 50% (at night) to 60% (during the day) of the time (K.F.A. Frumau, unpublished data).

The mature cloud forest was about 20–22 m tall, with *Ficus crassiuscula, Elaeagia auriculata, Weinmannia wercklei*, and several Myrtaceae being the most common tree species (Guariguata and Kattan, 2003) Tree ferns and palms (<4 m) dominated the understory. Epiphytes were abundant, reaching an estimated total biomass of 16.2 t ha^{-1} (Köhler *et al*., 2007) The *Pennisetum clandestinum* pasture site was dominated by herbaceous vegetation (mostly bracken) and had numerous cow trails that tended to follow the contours and which occupied, on average, about 30% of the watershed area.

Soils have developed in slightly to moderately weathered lahar deposits of the Quaternary Monteverde formation and in volcanic ashes deposited during a series of eruptive events since the late Miocene (Vargas, 2001) Soil profiles were multiple-layered and soil age increased with depth (Saigusa *et al*., 1987) The majority of the soils in the study area had similar upper horizons (A, A/B, Bw, and C) but their thickness varied considerably. They were classified as Udands, notably thaptic Hapludands, typic Hapludands, and typic Hydrudands (C. Tobón, unpublished data).

METHODOLOGY

General approach

To define the soil sampling scheme and the number of field sites required to evaluate the various soil physical and hydraulic properties, a detailed soil survey was carried out first in which the soils were classified at the great group level (Soil Survey Staff, 1996) Soil sampling and subsequent field and laboratory measurements were carried out between August 2003 and July 2004. The approximate effects of land use on soil properties were evaluated by comparing the specific properties of paired plots under undisturbed cloud forest and pasture.

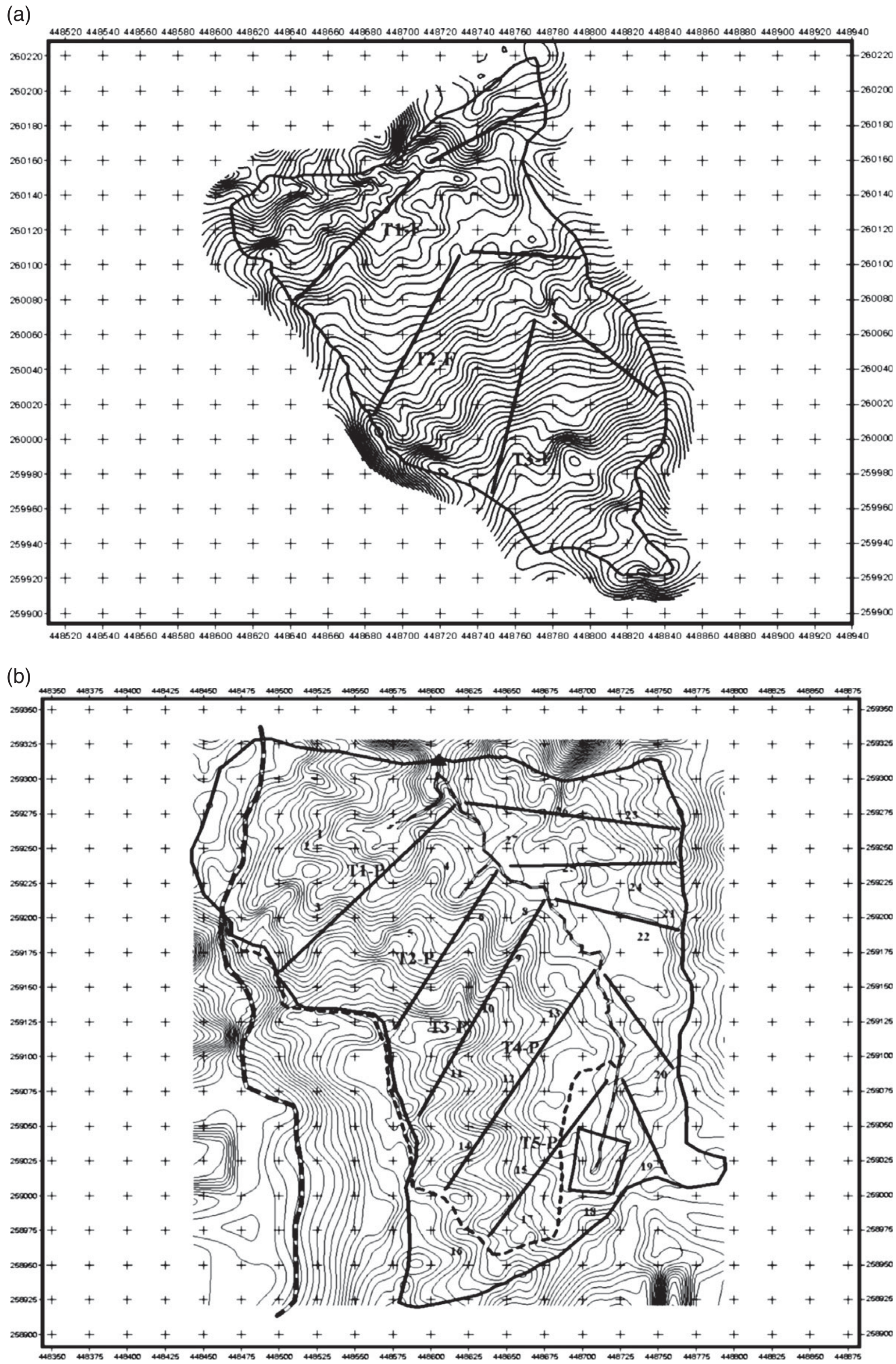

Figure 52.1. Soil sampling transects within (a) mature cloud forest (n = 19 sampling sites) and (b) grazed pasture (n = 27 sampling sites) near San Gerardo, Tilarán range, northern Costa Rica.

In the forest, three transects were laid out which ran from the nearest low ridge on either side of the stream to the corresponding water divides (Figure 52.1a), with a total of 19 sampling sites. In the pasture, five transects containing 27 sampling sites were selected (Figure 52.1b). At each site, soil samples for the laboratory determination of soil physical and hydraulic properties were collected from the different soil horizons (A, Bw, C, 2Ab, 2Bwb, 2C, A/Bb, 3Bwb, and 3C, where b denotes "buried") down to a depth of ~1.5 m (containing the majority of fine roots; Carvajal, 2003) In an attempt to relate and compare field and laboratory results, samples for laboratory analysis were taken close (0.5 m) to the points and at the same depths where the

Table 52.1 *Physical properties and soil organic matter of volcanic soils under cloud forest and grazed pasture near San Gerardo, Monteverde, northern Costa Rica. Average values, standard deviations, and the number of samples are presented, respectively, in each column*

Soil horizon	Texture		Bulk density (kg m^{-3})		Porosity		Soil organic matter (%)	
	Forest	Pasture	Forest	Pasture	Forest	Pasture	Forest	Pasture
A	Loam	Loam	406.5_19_100	527.6_165_123	0.85_0.01_100	0.79_0.07_123	16.2_2.3_10	7.8_3.6_14
A/B	Loam		484.2_21_100		0.81_0.01_34		13.2_3.4_10	9.6_3.3_14
Bw	Sandy loam	Sandy loam	611.8_38_49	626.7_117_103	0.77_0.02_49	0.76_0.04_103	7.8_1.7_10	9.2_2.9_14
C	Coarse to gravelly sand	Sand	712.4_21_64	689.6_100_81	0.73_0.02_64	0.73_0.04_81	2.7_0.78_10	4.6_2.1_14
Ab	Sandy clay	Sandy clay	639.0_159_47	633.7_153_97	0.77_0.06_47	0.76_0.06_97	9.9_3.1_5	10.2_2.9_5
Bw1	Loam	Sandy loam	667.4_191_54	639.5_110_77	0.74_0.06_54	0.75_0.04_77	3.8_0.98_5	6.2_1.8_5
2C1	Fine sand	Fine sand	811.1_80_48	889.5_158_91	0.69_0.04_48	0.66_0.06_91	2.45_0.56_5	2.5_0.87_5
A/Bb	Sandy clay	Sandy loam	693.7_132_66	647.8_141_90	0.73_0.05_66	0.75_0.05_90		
Bw2	Sandy clay	Sandy loam	723.4_121_51	773.9_136_69	0.72_0.05_51	0.70_0.05_69		
2C2	Loamy sand	Fine sand	778.0_76_50	791.0_177_84	0.70_0.05_50	0.70_0.07_84		

field measurements were carried out. All laboratory measurements were conducted at the Agricultural Research Center of the University of Costa Rica (CIA-UCR) in San José.

Soil texture, bulk density, soil organic matter, total porosity, and water retention characteristics

To determine soil bulk density, undisturbed samples were collected with stainless steel rings of 100 cm^3 at each site and depth. The samples were weighed and oven-dried at 105 °C for 48 hours. Bulk density was calculated as the ratio of the mass of the oven-dry sample to its bulk volume. In total, 456 samples were collected from the forest and 548 from the pasture.

For soil texture and particle size distribution, samples of 200 g were collected from each site and depth and subjected to a modified sieve–pipette sedimentation method using 1N NaOH as a dispergent and applying mechanical agitation (CIA-UCR, 1984) The USDA system was used for particle size classification and size limits (Soil Survey Staff, 1996) For soil organic matter (SOM) determination, three samples were collected from each soil horizon and at each forest and pasture site. Samples were mixed per soil horizon, and mean SOM content was determined for each soil horizon using the Walkey–Black method (Black, 1967) down to a depth of 100 cm (containing the first seven horizons; see Table 52.1). Total porosity, defined as total pore volume divided by sample volume, was evaluated from bulk density (kg m^{-3}) and average particle density (taken as 2600 kg m^{-3}; Baver, 1956)

For the characterization of soil water retention curves (WRCs), triplicate samples of 5 cm diameter and 1 cm height were collected at each site and depth. In total, 857 samples were collected, 392 in the forest and 465 in the pasture. Each core was wrapped in aluminum foil and placed in plastic bags to avoid disturbance during transportation to the laboratory. A complete retention curve was determined in order to allow evaluation of the amount of water lost at low suctions. First, the samples were weighed after reaching saturation on a sand box. For low levels of soil water suction (1, 3, 5, 10, and 20 kPa) a sand-box apparatus (Stakman and Van der Harst, 1969) was used. Samples were weighed after equilibration (normally within 48 hours). A pressure-plate apparatus was used to determine moisture contents associated with pressures of 30, 100, 300, 500, 1000, and 1500 kPa, with weighing after equilibration (Klute and Dirksen, 1986) The resulting experimental data pairs were fitted with the RECT 6.0 program (Riverside Salinity Laboratory, 2001) to obtain continuous mathematic descriptions for the WRCs. Because the shape of the WRCs indicated that many of the soils had a bimodal pore size distribution, the equations of Durner (1994) rather than those of Van Genuchten (1980) were used.

Infiltration and unsaturated hydraulic conductivity

An SMS, SW-080B tension infiltrometer (TI) was used to measure infiltration and derive the unsaturated hydraulic conductivity of surface soils through successive measurements at pressure heads of −15, −10, and −5 cm (Reynolds and Elrick, 1991) By adding a final measurement of steady-state infiltration at a tension close to zero, the saturated hydraulic conductivity could be estimated for comparison with the results obtained with other methods (see below). Generally, the time required to approach steady-state conditions varied from ~50 to 140 min. In total, 21 measurements were made in the cloud forest vs. 27 in the pasture, plus seven sites on cow trails. It was found that after the initial steady state was reached, the infiltration rate decreased for a while, after which it increased again to initial steady-state values. To better understand the infiltration dynamics and the underlying two- and three-dimensional soil water fluxes, soil water content was measured simultaneously at selected sites using small TDR probes (Campbell Scientific, 10-cm length) placed underneath the infiltrometer disk, and at specific depths (i.e. at 5, 10, and 15 cm in the A- and Bw-horizons, and at 22 and 25 cm in the top and middle part of the coarse C-horizon). Infiltration rate and soil water content were measured again after 1 week (Vogeler *et al*., 1996; Wang *et al*., 1998).

Saturated hydraulic conductivity

Values of saturated hydraulic conductivity (K_s) appear to be influenced by the method used. In addition, values are highly sensitive to sample size, flow geometry and sample collection procedures (Bouma, 1983; Reynolds and Elrick, 1991; Stolte *et al*., 1994; Davis *et al*., 1999) Therefore, K_s was characterized using both field and laboratory methods and differently sized samples, viz. (i) the constant-head method using small soil cores (7.5. × 7.5 cm) in the laboratory (Klute and Dirksen, 1986), (ii) field tension infiltrometry (with 20-cm diameter disk), and (iii) the constant-head well permeameter method in uncased boreholes of 6-cm diameter (Guelph permeameter; Reynolds and Elrick, 1985, 1991).

Soil macropores, when filled with water, permit rapid turbulent flow (Beven and Germann, 1982) which may disturb core samples during laboratory experiments. Therefore, a fine nylon cloth (250-μm mesh width) was placed at the base of the core samples to avoid loss of material. In the tension infiltrometer approach the algebraic approximation for steady-state unconfined saturated infiltration rates as proposed by Wooding (1968) was used to calculate K_s from measured infiltration. Determination of field K_s in uncased boreholes involves the use of analytical solutions which are based on the assumption that the soil is homogenous throughout the profile (Reynolds and Elrick, 1985; Xiang, 1994) This condition was not met in the layered soils under consideration. According to Xiang *et al*. (1997) it is important to estimate the conductivity for each soil layer separately in such cases. Therefore, measurements of steady-state rates were made at two different depths, usually at 0.1 m and at 0.2–0.3 m (representing the A- and Bw-horizons and the C-horizon, respectively) but occasionally at greater depth (0.5 m) if required by horizon distribution. Successive measurements were made in separate boreholes placed at a distance of *c*. 0.5 m. In this way, effects of soil heterogeneity were minimized while avoiding side effects of the previous measurement.

Statistical analysis

In general, the experimental design for the statistical analysis was randomized per soil horizon. Soil properties were compared between soil depths and sites within the forest or pasture, and between forest and pasture, on the basis of geometric means, minimum and maximum values, the coefficient of variation, and the Pearson correlation coefficient. Data were analyzed using analysis of variance (ANOVA). Differences were considered significant at 0.05 probability level. For the spatial distribution of K_s, a log-normal statistical distribution was assumed, a common procedure for *in situ* and undisturbed core measurements (Warrick and Nielsen, 1980; Reynolds *et al*., 2000; see also Zimmermann and Elsenbeer (2008) for in-depth discussion). Therefore, the data were log-transformed (ln) prior to analysis and the antilog taken before interpretation of the results. Correlations between results obtained with different methods were conducted on a site-to-site basis within each land-cover type and soil layer using the ln-transformed K_s values.

RESULTS AND DISCUSSION

Soil texture

Table 52.1 gives the average values and standard deviations for texture, bulk density, and porosity of each soil horizon down to a depth of 1.5 m. In both forest and pasture, sandy textures were predominant throughout the soil profiles and along transects. The uppermost C-horizon consisted mainly of sandy and gravelly particles of about 2 mm (lapilli) that were deposited throughout the study area during the last major volcanic event (Vargas, 2001) Differences in particle size distribution of corresponding horizons beneath forest and pasture were not significant ($p < 0.05$). However, in the forest, the first C-horizon contained a larger amount of organic material than in the pasture. This organic matter is thought to have derived from the surface layer (Ao) and to have accumulated at the boundary between the highly permeable C-horizon and the denser buried 2Ab-horizon below, presumably due to the sudden decrease in hydraulic conductivity (cf. Table 52.2 below). In the pasture, the sandy C-horizon material was partly mixed with finer material eluviated from the upper Ap-horizon, producing a more loamy overall texture (Table 52.1).

The predominantly sandy texture of the soils is mainly due to the presence of the volcanic ashes, and partly the result of wetting and drying cycles producing the formation of micro-aggregates (Kubota, 1976; Shoji *et al*., 1993; cf. Schrumpf *et al*., this volume). Moreover, non-crystalline materials present in volcanic soils act as cementing agents and hamper complete dispersion of mineral particles during textural analysis (Ping *et al*., 1988), typically resulting in an underestimation of clay content.

Bulk density

Bulk densities were very low throughout the profiles, except for the gravelly C-horizon which showed the highest values (Table 52.1). Low density is a special feature of volcanic soils and related to their highly porous structure, reflecting an abundance of non-crystalline materials and organic matter (Shoji *et al*., 1993). The bulk density values obtained for the San Gerardo cloud forest soils are among the lowest reported for volcanic soils, but they are comparable to other young volcanic soils (Shoji *et al*., 1993), with similar amounts of organic matter (Table 52.1). Comparing bulk density values within the same profile, there were no significant differences ($p < 0.05$) between different soil layers, both under forest and pasture, except for the top-soil (A- and A/B-horizons) which exhibited significantly lower values under both land-cover types. However, the difference was considerably less in the case of pasture (Table 52.1).

Bulk densities of each soil layer were also compared (horizontally) between sampling sites (separately for forest and pasture). In the forest, surface soil layers showed no significant differences ($p < 0.05$) between sites but there were large differences for sub-surface layers as evidenced by their large standard deviations (Table 52.1). This reflects differences in parent material which varied from sandier volcanic ashes, to lahar deposits producing more clayey sub-soil horizons.

As for the pasture, differences in soil bulk density between sites were much larger, notably for the first 25 cm (Table 52.1). This was mainly due to the deviating values obtained for profiles positioned on slopes where the original A-horizon had become mixed with the underlying A/B- or Bw-horizons. At some sites, the entire top-soil had become mixed with C-horizon material, producing a new top-layer with a coarser texture (sandy loam), possibly due to soil disturbance caused by cattle activity. Moreover, there was a trend toward increasing (top-soil) density values going from the ridges to riparian footslopes. This seems related to erosion of soil particles from the upper parts of the slopes and from cow trails by overland flow (C. Tobón, unpublished data) followed by deposition on the gentler footslopes (Harden, 1992) Comparing soil bulk densities of the first 25 cm below cow trails and non-trail surfaces showed even larger differences ($p < 0.01$). The regular passage of cows and people along the paths produced much higher density values (Figure 52.2a), as has also been found for coarser-textured soils elsewhere (Patel and Singh, 1981; Harden, 1992; Ziegler *et al*., 2004).

Comparisons of mean bulk density values between forest and pasture show these to differ significantly ($p < 0.05$) in the first 25 cm, with values being larger in the pasture (Figure 52.2a) However, for deeper layers, densities did not differ much between treatments (Pearson r = 0.89, $p = 0.0014$) (Figure 52.2). Spatial variability in C-horizon bulk density under pasture was much higher than under forest (Table 52.1), although the degree of mixing varied between sites.

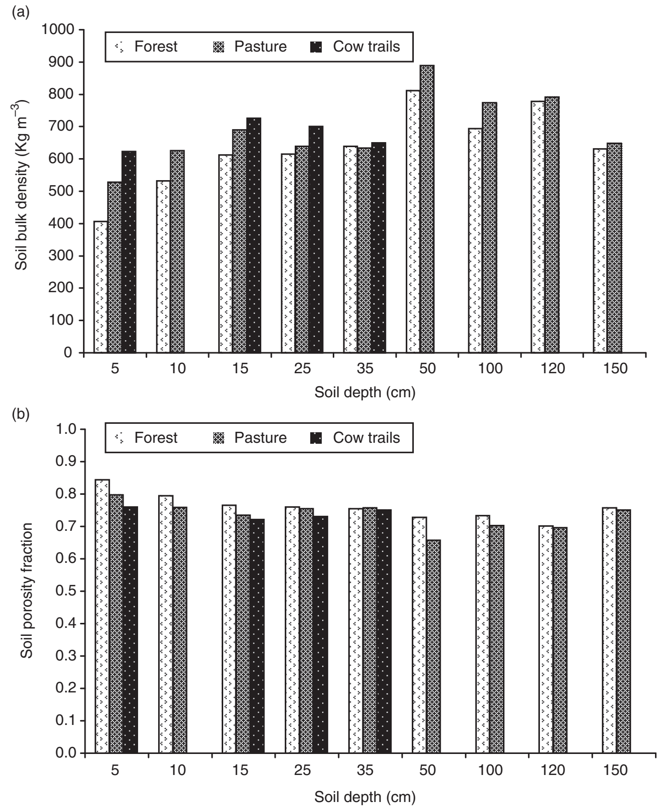

Figure 52.2. (a) Bulk density (kg m^{-3}) and (b) porosity (fraction) of the different soil horizons beneath mature cloud forest and grazed pasture sites near San Gerardo, Tilarán range, northern Costa Rica.

Porosity

In line with the low density and well-developed structure of the soils, total porosity values were very high at 0.69–0.85 in the forest and 0.66–0.79 in the pasture (Table 52.1; cf. Furuhata and Hayashi, 1980) For most soil horizons significant differences occurred between forest and pasture ($p < 0.05$), whereas within the pasture significant differences ($p < 0.05$) were observed between top-soil porosities associated with cow trails and non-trail surfaces (Figure 52.2b).

The average 30% increase in A-horizon bulk density in the pasture compared to the forest produced an average reduction in top-soil porosity of only 7% (Table 52.1). As will be shown below, this change was primarily effected by a reduction in pore volume associated with macropores.

Water retention characteristics

The water retention curves (WRCs) for the respective layers of the forest soil showed broadly similar tendencies indicating broadly comparable soil properties (Figure 52.3a) Similar trends were observed for the pasture soils (Figure 52.3b), except below the cow trails which showed completely different behavior

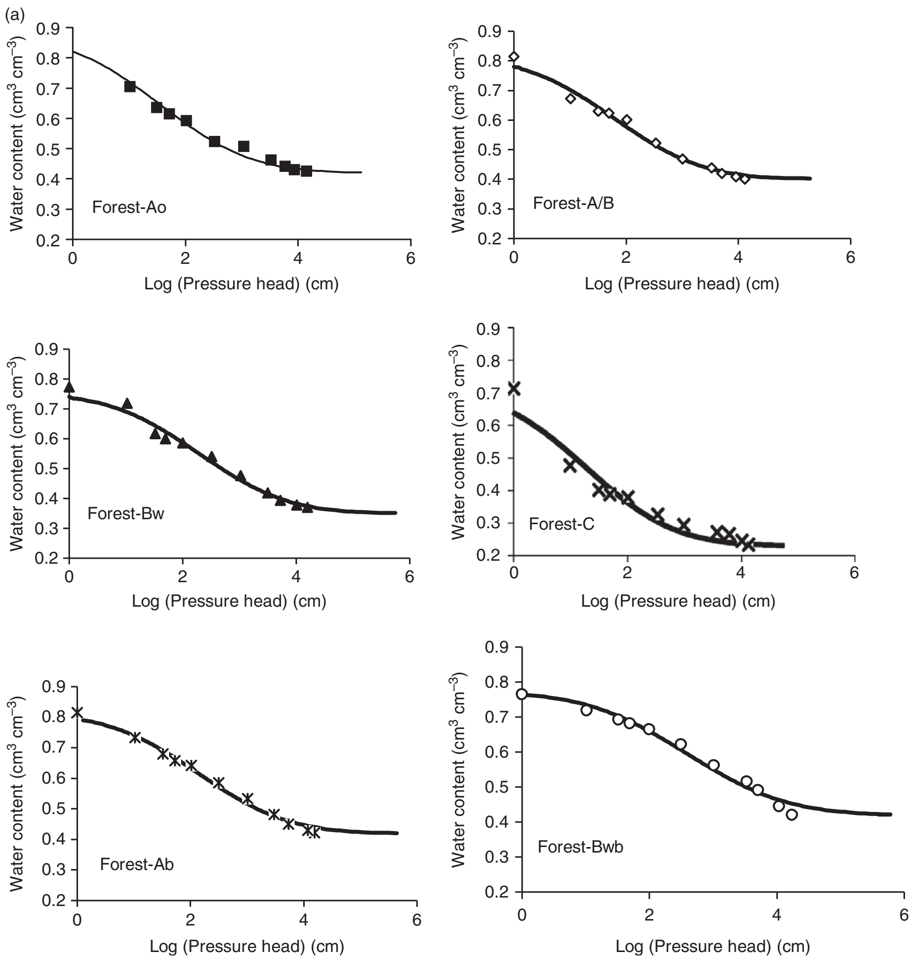

Figure 52.3. Measured and fitted water retention curves for soils under (a) mature cloud forest, (b) grazed pasture, and (c) compacted cow trails near San Gerardo, Tilarán range, northern Costa Rica.

(rather resembling that of clayey soils; Figure 52.3c) because of their compacted nature. Most WRCs resembled the curves normally derived for coarse-textured soils, despite the very low bulk densities of the present soils. According to Luxmoore (1981), macropores empty at pressures between 0 and 0.3 kPa and mesopores between 0.3 and 30 kPa. Thus, a slight trend was discernible toward increased soil water retention with depth, reflecting the larger numbers of mesopores (2–200 μm) and micropores (<2 μm) at depth compared to a dominance of macropores (>200 μm) and micropores near the surface (Luxmoore, 1981) As a result, water content close to saturation was larger in surface horizons than in deeper layers (Figure 52.3a,b).

Generally, plant-available water (*PAW*) is defined as the amount of water held between field capacity and permanent wilting point, estimated as the difference in water contents between soil matric potentials of 33 and 1500 kPa (Marshall *et al.*, 1996). In the forest, values of *PAW* defined in this way, ranged from 0.09 to 0.17 $cm^3\ cm^{-3}$ in the upper layers and from 0.11 to 0.20 $cm^3\ cm^{-3}$ in deeper layers. In the pasture, values varied from 0.13–0.19 $cm^3\ cm^{-3}$ in top-soil vs. 0.12–0.18 in deeper layers. Whilst a suction value of 1500 kPa is widely accepted to characterize the permanent wilting point for many plants growing in a wide range of soils, the generally used value for field capacity (33 kPa) is not appropriate for volcanic soils. Instead, for Andisols formed under humid climates, a value of 10 kPa is more suitable (Saigussa *et al.*, 1987) Taking 10 kPa to denote field capacity, *PAW* values increased by up to 80% for the forest soils and up to 35% for the pasture soils, with the largest increases noted for top-soils.

The studied soils appeared to lose a large part of their water already at comparatively low suctions (*c*. 36% at suctions <33 kPa). At the same time, the shape of the WRCs indicated that most soil horizons were capable of retaining large amounts of water at wilting point (1500 kPa). Indeed, the present soils retained roughly twice the amount of water held by highly clayey

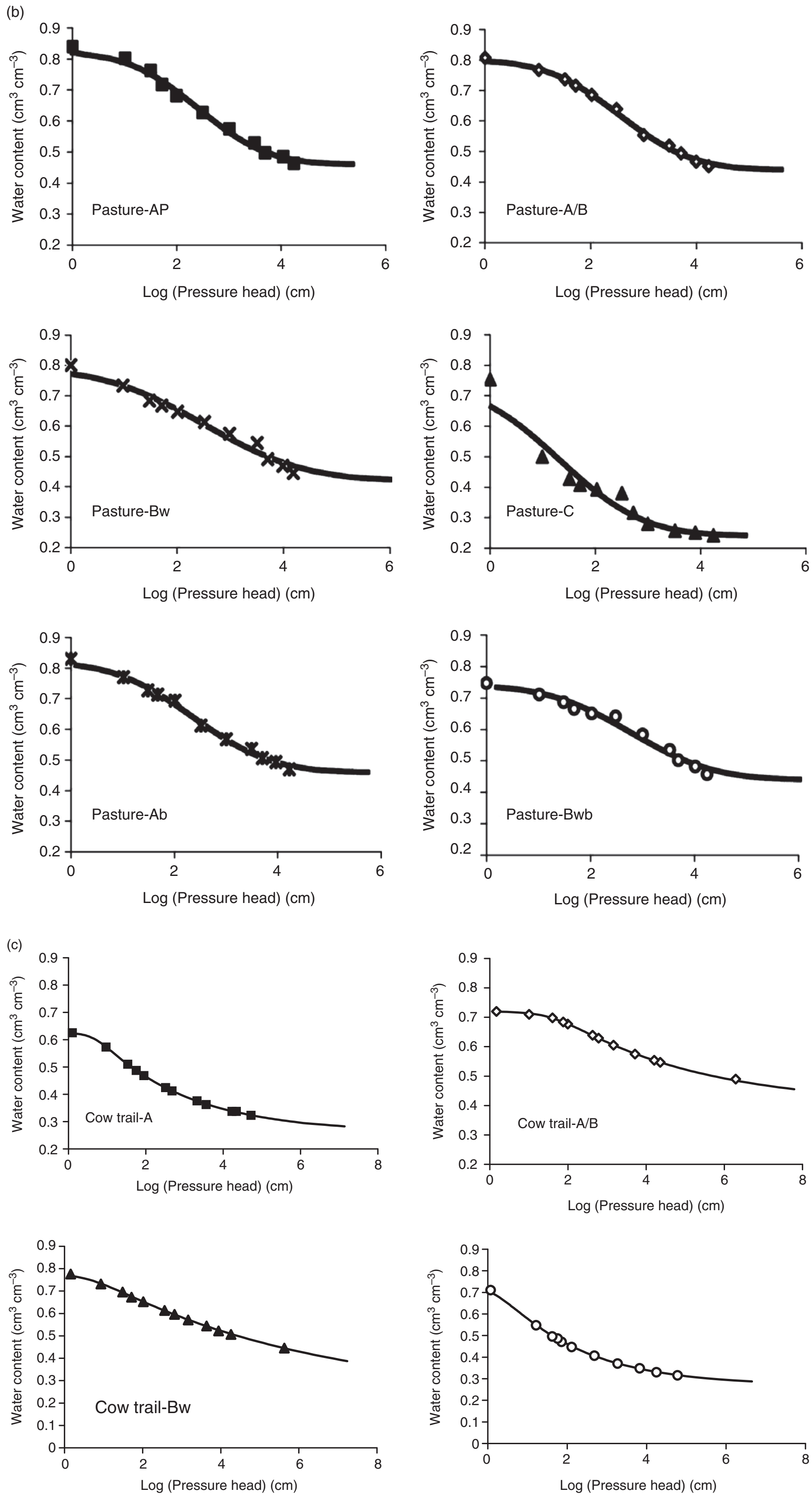

Figure 52.3. (*cont.*)

soils in the Amazon (Tobón, 1999; Tomasella *et al*., 2000) This high residual water retention of volcanic soils has been attributed to their high concentrations of non-crystalline materials (notably allophanic clays) intimately mixed with organic matter (Saigusa *et al*., 1987; Ito *et al*., 1991).

The WRCs suggest considerable differences in water retention capacity between the upper soil layers (A-, A/B-, and Bw-horizons) and the C- and Bwb-horizons below, particularly in the cloud forest. Whilst curves for the upper horizons clearly show a bimodal pore size distribution (dominance of macro- and micropores), in deeper horizons this trend is less clear, implying fewer macropores. The curves for soil layers below 1 m depth generally (but not always) showed only a small reduction in soil water content with increasing suction (implying a dominance of micropores), as is normally found for clayey soils with a low concentration of roots and low microbial activity.

Comparing WRCs for forest and pasture soils shows that although water availability is higher in the forest for most layers, the water retained at 33 kPa is always higher in the pasture. This is thought to reflect changes in the relative abundances of macropores and their replacement by either meso- or micropores. The disappearance of the forest soil's dual pore size distribution in the upper horizons after conversion to grazed pasture indicates the loss of a large proportion of the macro- and mesopores, particularly on the cow trails (Figure 52.3c). Whilst the WRCs for the layers below the first C-horizon were rather similar for forest and pasture, those for the C-horizons themselves were rather different (Figure 52.3ab). In the forest, the coarse-textured C-horizon showed very fast water release even at low suctions (<1 kPa), whilst for the more loamy C-horizon of the pasture the release was more gradual. In addition, the pasture C-horizon retained a very large amount of water was at 1500 kPa (Figure 52.3b). As indicated previously, the pasture C-layer has received organic and fine material from the top-soil, which has rendered its texture more loamy.

Infiltration

Figure 52.4 shows representative field infiltration curves obtained for the cloud forest and the pasture. Average infiltration rates at the beginning of infiltration were only twice the final ponded values, suggesting that macropores dominated water flow under saturated conditions. The decrease in infiltrability with time seems to occur quite fast (Figure 52.4), particularly after conversion to pasture where the gradual disappearance of macropores due to compaction, and the accumulation of a dense mass of grass roots tended to reduce the high initial rates typically associated with the undisturbed situation (Zimmermann and Elsenbeer, 2008).

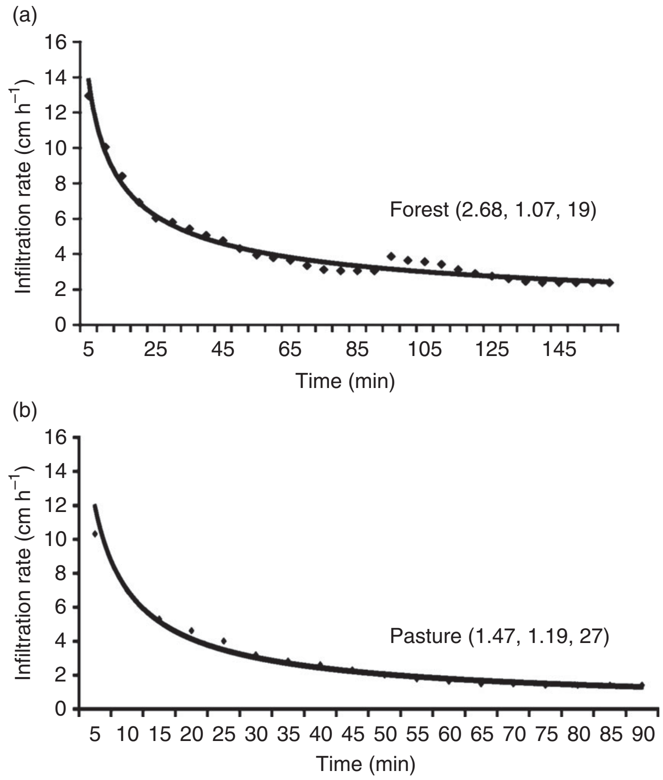

Figure 52.4. Infiltration curves for top-soils under (a) mature cloud forest and (b) grazed pasture near San Gerardo, Tilarán range, northern Costa Rica. Numbers shown are final steady-state rate, standard error, and number of measurements, respectively.

Values of infiltration rate at different locations within the forest (even between trails and non-trails) were not significantly different ($p < 0.05$), but in the pasture (also on the cow trails), spatial variability was high. Average surface infiltration rate in the forest (f_c) was 2.68 ± 1.07 cm h^{-1} vs. 1.47 ± 1.19 cm h^{-1} in the pasture (non-trail surfaces) and 0.87 ± 0.34 cm h^{-1} on the cow trails. After inclusion of the values for cow trails in the pasture, differences between forest and pasture became significant ($p < 0.05$). The lower infiltration rates on the cow paths resulted in increased frequency and amounts of overland flow, both on the trails themselves and in adjacent downslope grassy areas (C. Tobón, unpublished data).

In the pasture, patterns of infiltration were systematically related to slope position; rates were relatively high on the upper and middle parts of the slopes, and lowest on the footslopes and in concave areas close to the stream where soils had higher bulk densities, and fewer macropores according to their WRCs (cf. Harden and Scruggs, 2003).

It was expected that final infiltration rates f_c would approach saturated hydraulic conductivity K_s. However, f_c was typically lower than corresponding K_s (see Table 52.2 below), which may be related to the development of a perched water table above the gravelly C-horizon during the infiltration experiments. As

Table 52.2 *Values of geometric means, maxima, minima, coefficient of variation (CV), and number of samples (*n*) of saturated hydraulic conductivity (cm day^{-1}) of different layers of volcanic soils under cloud forest, grazed pasture, and cow trails near San Gerardo, Monteverde, northern Costa Rica*

Soil	Guelph permeameter					Tension infiltrometer					Small-cores method				
horizon	Mean	Max.	Min.	CV%	*n*	Mean	Max.	Min.	CV%	*n*	Mean	Max.	Min.	CV%	*n*
Forest															
A	71.4	183.4	34.5	54.4	16	37.1	62.3	9.2	10.4	21	47.2	90.9	15.9	53.4	30
Bw	18.1	21.4	15.7	16.0	5						15.7	20.7	9.6	23.7	19
C	1204	2563	658.1	83.8	5						3171	4827	2235	31.2	12
2Ab	37.6	57	28.4	33.4	5						29.6	63.4	15.9	60.8	12
2Bwb											13.3	51.0	4.5	146	7
Pasture															
A	24.3	48.4	2.9	88.5	17	17.7	38.2	2.2	10.5	27	16.9	37.4	7.2	53.6	25
Bw	29.5	56.7	17.3	51.0	5						24.1	41.1	6.5	37.5	14
C	580	1025	193	62.6	5						1414	1761	1143	15.1	14
2Ab	14.2	45.7	4.6	127.1	5						14.1	24.0	7.4	29.5	11
2Bwb											8.5	15.9	2.3	52.7	9
Cow trails															
A-Bw						3.2	8.3	1.3	2.5	7	3.3	105.4	0.3	1401	9

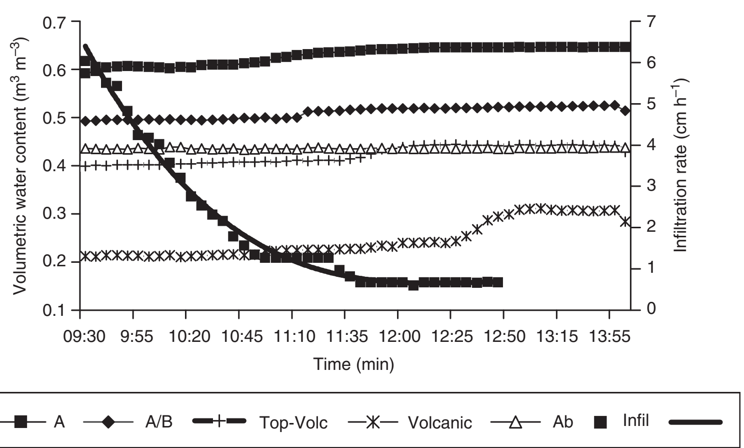

Figure 52.5. Trends in infiltration rate as a function of soil water content in a layered volcanic soil beneath mature cloud forest, San Gerardo, Tilarán range, northern Costa Rica.

illustrated in Figure 52.5, soil water content in the top (A) layer changed moderately after the start of infiltration, while underneath this layer (A/B) a response was only observed later (although before attaining steady-state conditions). This indicates that water moved from the less permeable A- and Bw-horizons to the more permeable C-horizon below. Once the wetting front reached the top of this coarse layer, ponding occurred because of the poor contact between the two layers. Only after saturation was reached at the top of the coarse layer, water began to move through it, albeit at a rate controlled by the conductivity of the less permeable horizons above it, and until the wetting front reached the 2Ab- or 2A/Bb-horizons found below the C-horizon. These deeper layers had considerably lower permeabilities than either the gravelly C-horizon or the top-layers (Table 52.2 below).

Saturated hydraulic conductivity

Table 52.2 presents average, maximum, and minimum values as well as the coefficient of variation of K_s for the different profiles

Table 52.3. *Site-wise Pearson correlation coefficient (r) and statistical significance* (p-*value*) *of differences in saturated hydraulic conductivity as obtained using three different methods: Tension infiltrometer (TI), Guelph permeameter (GP), and intact soil cores (SC). The analysis was conducted using ln-transformed data from cloud forest and grazed pasture near San Gerardo, Monteverde, Costa Rica*

Land Cover	Methods compared	r-value	p-value
Cloud forest	GP vs. TI	0.2876	0.3768
	GP vs. SC	0.2121	0.4563
	TI vs. SC	0.0835	0.7821
Pasture	GP vs. TI	0.4722	0.2457
	GP vs. SC	0.3864	0.3123
	TI vs. SC	0.2156	0.4465

as obtained with the three methods used (i.e. constant-head permeametry in the laboratory, tension infiltrometry, and Guelph permeameter in the field). Significant differences were found between the respective soil layers (vertical comparison) per land cover, the differences being largest between the surface layers (A–Bw) and the C-horizon, both under forest and under pasture (Table 52.2). Values of K_s under forest were generally significantly higher ($p < 0.05$) than beneath pasture for corresponding soil horizons (horizontal comparison) and for all three methods. The higher values for the forest sites are primarily related to the presence of large macropores and a well-developed soil structure. K_s reached its maximum values in the coarse-textured upper C-horizon but decreased rapidly in the layers underneath.

Comparing corresponding soil layers for the two land-cover types, spatial variability in K_s was much lower in the forest soils regardless of the method used, in agreement with the pattern described earlier for soil bulk density (Table 52.1). Spatial variability in pasture K_s was increased further when including the cow trails (Table 52.2). However, although spatial variability was larger in the pasture, the CV of K_s in the C-layer was larger in the forest, possibly because of spatial variations in amounts of illuviated organic matter. The high CV values obtained with the small-core method for the 2Ab- and 2Bwb-layers underneath the C-horizon mainly reflect variability in soil texture resulting, in turn, from differences in parent material (Table 52.1).

Top-soil K_s values as measured with the Guelph permeameter and on small cores were consistently lower close to the stream (narrow riparian area) than at upslope locations, particularly in the pasture. This pattern is the inverse of that found earlier for soil bulk density and caused a decrease in the time to ponding in a downslope direction, such that overland flow occurrence was increased even during small rainfall events (cf. Harden and Scruggs, 2003).

Table 52.2 also shows that K_s values differed between methods. Generally, for all soil layers and land covers, values of field K_s obtained with the Guelph permeameter were the highest, except for the C-layer, for which the small-cores method gave higher values. Although comparisons between different techniques for measuring K_s suffer from the lack of independent benchmark values, an attempt is made below to explain some of the observed contrasts. The higher K_s values found generally with the Guelph permeameter can be explained by the larger sample size (i.e. vertical area) employed by this method, thereby allowing more representative inclusion of large macropores (compared to the small-cores method; Davis *et al*., 1999) and multidimensional flow (compared to the one-dimensional flow imposed by tension infiltrometry; Reynolds *et al*., 1985, 2000). Also, although most 15 cm deep auger holes included only the A-horizon, at some pasture sites they also included the A/B- or Bw-horizon. Because K_s values of the latter were generally higher than that of the A-horizon (Table 52.2), their inclusion would positively influence the result.

Tension infiltrometry yielded significantly ($p < 0.05$) lower mean, maximum, and minimum values of top-soil K_s values compared to those obtained with the Guelph permeameter or the small-cores method (with the exception of the pasture in the latter case). This may be related to one or both of the following reasons: while smoothing the soil surface to install the ceramic plate prior to the measurement, even if done very carefully, surface soil pores may be destroyed or become partially closed (Schwärzel and Punzel, 2007). Alternatively, a lower pressure head than imposed may develop in part of the contact sand between the infiltrometer disk and the underlying soil, which produces a decrease in flow through macropores relative to what would have occurred without the presence of the contact sand (Reynolds *et al*., 2000).

In the sandy to gravelly C-layer, values of K_s obtained with the small-cores method were 2.4–2.6 times higher than those produced by the Guelph permeameter, whereas values were similar between methods for the other soil horizons (Table 52.2). This may reflect rapid preferential flow through the large macropores of this coarse layer which may have extended along the entire length of the sample cores. Moreover, it was observed that at the beginning of the experiments water passing through the samples flushed out organic matter and other fine material, thereby leaving only gravel and coarse sand. Indeed, steady-state K_s values were higher at the end of the experiments although the samples were completely saturated from the start.

The lower K_s values obtained with the small-cores method for the other, less coarse layers might theoretically be explained by compaction during coring or simply to inadequate sample size (Bouma, 1983; Davis *et al*., 1999; Reynolds *et al*., 2000; Wu *et al*., 1992). However, compaction of the present samples is unlikely because of their sandy loam texture and careful

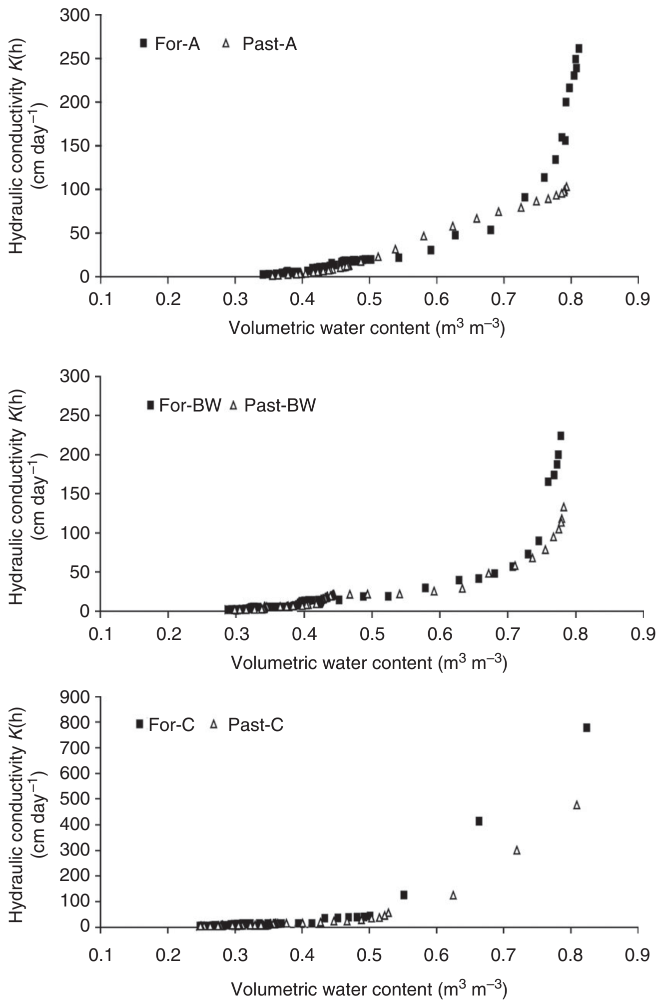

Figure 52.6. Relationships between unsaturated hydraulic conductivity (cm h^{-1}) and volumetric water content (cm^3 cm^{-3}) for three main horizons within layered volcanic soils beneath mature cloud forest and grazed pasture near San Gerardo, Tilarán range, northern Costa Rica.

collection and transport. Moreover, large samples are more difficult to collect and transport without disturbance. Rather, the present results can be explained by the fact that the soils had visible pores in all directions that were easily closed off by the core walls upon sampling, thereby restricting preferential flow.

The value of K_s measured in the field includes any effects of entrapped air and may be twice as small as the true saturated value (Stephens *et al*., 1983; Reynolds and Elrick, 1985). However, this is unlikely to be the case in the present soils which remained wet throughout the study period, thereby impeding air entry to most soil pores. Therefore, the values obtained with the Guelph permeameter are considered the most representative.

Unsaturated hydraulic conductivity

Figure 52.6 presents the unsaturated hydraulic conductivity K_h as a function of volumetric water content (θ) for the three main soil layers (A, Bw and C) in the forest and pasture. As found earlier for the WRCs (Figure 52.3), the conductivities for the upper two soil layers in the forest show the dominant influence of macropores, in that conductivity at high water content ($\theta > 0.75$) decreased rapidly for only small decreases in water content. At low water contents ($\theta < 0.40$) the changes in K_h were very small. Conversely, K_h in the sandy C-layer decreased to a very low value already at a water content of about 0.50 cm^3 cm^{-3} (Figure 52.6). Thus, whilst K_s of the C-horizon greatly exceeded that of the A- or Bw-horizon (Table 52.2), the opposite was true for K_h at relatively minor suction values (10–30 kPa) corresponding to θ values of 0.47–0.42.

Relationships of K_h and θ differed considerably between forest and pasture, particularly for the A-horizon and to a lesser extent the C-horizon (Figure 52.6). For example, at a suction of 10 kPa (corresponding to a θ of 0.59 cm^3 cm^{-3}), K_h in the A-horizon of the forest was about three orders of magnitude higher than in the pasture, while in the Ab-horizon below the C-layer it was still twice as high (C. Tobón, unpublished data). However, at relatively low water contents ($\theta < 0.45$ cm^3 cm^{-3}) K_h was similar in forest and pasture soils, except for the cow trails where K_h was very low throughout the moisture range (not shown). Such contrasts are associated with the reduction in the number of macropores in the pasture and their replacement by meso- and micropores in the top layers (Ap and Bw). The present results agree with the observations of Buytaert (2004) and Podwojewski *et al*. (2008) on volcanic soils in Ecuador and illustrate the fragility of these soils.

CONCLUSIONS

The present results demonstrate that the soil physical and hydraulic properties of volcanic soils under old-growth montane cloud forest are rather unique in that they have very low bulk densities throughout the soil profile, high porosity (provided mainly by macropores), high water retention at 1500 kPa, high plant-water availability (between suctions of 10 and 1500 kPa) and high infiltration rate and hydraulic conductivity. Conversion to grazed pasture produced important changes in these properties, particularly in the top 25 cm. Specifically, top-soil infiltration and hydraulic conductivity were strongly reduced, as was soil porosity, whereas bulk densities were increased. These changes reflect changes in pore size distributions, notably a decrease in the number of macropores and their replacement by meso- and micropores. The greatest changes were observed for the soils underneath cow trails in the pasture, where infiltration was greatly diminished and runoff production increased.

The large difference in K_s of the top-soil and the C-horizon under forest is likely to result in a stormflow regime dominated by fast lateral sub-surface flow (Bonell, 2005) However, on the steep and upper slopes of the pasture watershed, the gravelly C-horizon had become mixed with material from the layers above, resulting in a more loamy texture, slightly lower bulk density and considerably reduced K_s, as well as in improved water retention compared to the original C-horizon. Thus, contrasts between the altered C-horizon and the layers above and below were less pronounced and flows to deeper layers more continuous. This may lead, in turn, to a less responsive lateral sub-surface flow regime.

The results of this study indicate that land cover had a major effect on key soil properties, and consequently on water movement in these layered volcanic soils. As such, parameterization of soil hydraulic properties in watershed models based on soil type or texture alone is likely to lead to poor predictions of the hydrologic impact of cloud forest conversion to pasture.

ACKNOWLEDGEMENTS

Financial support for this study by the Forestry Research Programme of the Department for International Development (DFID) of the UK (project R7991) is gratefully acknowledged. The views expressed here are not necessarily those of DFID. Eulogio Jimenez and the owners of the forest and pasture catchments are thanked for their permission to use the sites.

REFERENCES

Baver, L.D. (1956). *Soil Physics*. New York: John Wiley.

Beven, K., and P. Germann (1982). Macropores and water flow in soils. *Water Resources Research* **18**: 1311–1325.

Black, C.E. (1967). *Method of Soil Analysis*, American Society of Agronomists Monograph No. 9. Madison, WI: American Society of Agronomists.

Bogner, C., S. Engelhardt, J. Zeilinger, and B. Huwe (2008). Visualization and analysis of flow patterns and water flow simulations in disturbed and undisturbed tropical soils. In *Gradients in a Tropical Mountain Ecosystem of Ecuador*, eds. E. Beck, J. Bendix, I. Kottke, F. Makeschin, and R. Mosandl, pp. 403–412. Berlin: Springer-Verlag.

Bonell, M. (2005). Runoff generation in tropical forests. In *Forests, Water and People in the Humid Tropics*, eds. M. Bonell and L.A. Bruijnzeel, pp. 314–406. Cambridge, UK: Cambridge University Press.

Bouma, J. (1983). Use of soil survey data to select measurement techniques for hydraulic conductivity. *Agricultural Water Management* **6**: 177–190.

Buytaert, W. (2004). The properties of the soils of the south Ecuadorian *páramo* and the impact of land use changes on their hydrology. Ph.D. thesis, Catholic University Leuven, Leuven, Belgium.

Carvajal, A. (2003). Distribución de raíces finas en suelos del bosque nuboso y pastos, en Monteverde, Costa Rica. B.Sc. thesis, Instituto Tecnológico de Costa Rica, Cartago, Costa Rica.

Centro de Investigaciones Agronómicas (CIA-UCR) (1984). *Manual de laboratorio de edafología*. San José, Costa Rica: CIA-UCR.

Clark, K.L., R.O. Lawton, and P.R. Butler (2000). The physical environment. In *Monteverde: Ecology and Conservation of a Tropical Cloud Forest*, eds. N.M. Nadkarni and N.T. Wheelwright, pp. 15–34. Oxford, UK: Oxford University Press.

Davis, S.H., R.A. Vertéis, and R.P. Silberstein (1999). The sensitivity of a catchment model to soil hydraulic properties obtained by using different measuring techniques. *Hydrological Processes* **13**: 677–688.

Duisberg-Waldenberg, P. (1980). Erosión y conservación de suelos. In *Estudio ecológico integral de las zonas de afectación del Proyecto Arenal*, Centro Científica Tropical, 2–1–2–65. San José, Costa Rica: Centro Científica Tropical.

Durner, W. (1994). Hydraulic conductivity estimation for soils with heterogeneous pore structure. *Water Resources Research* **30**: 211–223.

Edwards, K.A. (1979). The water balance of the Mbeya experimental catchments. *East African Agricultural and Forestry Journal* **1**: 231–247.

Frumau, K.F.A., L.A. Bruijnzeel, and C. Tobón (2006). Measurement of precipitation in montane tropical catchments: comparative performance of conventional, spherical and "potential" rain gages. In *Forest and Water in a Changing Environment*, eds. S.R. Liu, G. Sun, and P.S. Sun, pp. 104–108. Vienna: IUFRO, and Beijing: Chinese Academy of Forestry.

Furuhata, A., and S. Hayashi (1980). Relation between soil structure and soil pore composition: case of volcanogenous soils in Tokachi district. *Research Bulletin o the Hokkaido National Agricultural Experimental Station* **126**: 53–58 (in Japanese, with English summary).

García-Santos, G. (2007). *An ecohydrological and soils study in a montane cloud forest in the National Park of Garajonay, La Gomera (Canary Islands, Spain)*. Ph.D. thesis, VU University Amsterdam, Amsterdam, the Netherlands. Also available at www.falw.vu.nl/nl/onderzoek/earth-sciences/geo-environmental-science-and-hydrology/hydrology-dissertations/index.asp.

Giambelluca, T.W., R.E. Martin, G.P. Asner, *et al*. (2009). Evapotranspiration and energy balance of native wet montane cloud forest in Hawai'i. *Agricultural and Forest Meteorology* **149**: 230–243.

Grip, H., J.M. Fritsch, and L.A. Bruijnzeel (2005). Soil and water impacts during forest conversion and stabilization to new land use. In *Forests, Water and People in the Humid Tropics*, eds. M. Bonell and L.A. Bruijnzeel, pp. 563–589. Cambridge, UK: Cambridge University Press.

Guariguata, M.R., and G.H. Kattan (2003). *Ecología y conservación de bosques Neotropicales*. San José, Costa Rica: Editorial Tecnológica de Costa Rica.

Hafkenscheid, R.L.L.J., L.A. Bruijnzeel, R.A.M. deJeu, and N.J. Bink (2002). Water budgets of two upper montane rain forests of contrasting stature in the Blue Mountains, Jamaica. In *Proceedings of the 2 International Colloquium on Hydrology and Water Management in the Humid Tropics*, ed. J.S. Gladwell, pp. 399–424. Paris: IHP-UNESCO, and Panama City: CATHALAC.

Hamilton, L.S., J.O. Juvik, and F.N. Scatena (eds.) (1995). *Tropical Montane Cloud Forests*. New York: Springer-Verlag.

Harden, C.P. (1992). Incorporating roads and footpaths in watershed-scale hydrologic and soil erosion models. *Physical Geography* **13**: 368–385.

Harden, C.P., and P.D. Scruggs (2003). Infiltration on mountain slopes: a comparison of three environments. *Geomorphology* **55**: 5–24.

Hermann, R. (1971). Die zeitlichen Anderung der Wasserbindung im Boden unter verschiedenen Vegetationsformationen der Hohenstufen eines tropischen Hochgebirges (Sierra Nevada de Santa Marta, Kolumbien). *Erdkunde* **25**: 90–102.

Holwerda, F. (2005). Water and energy budgets of rain forests along an elevation gradient under maritime tropical conditions. Ph.D. thesis, VU University Amsterdam, Amsterdam, the Netherlands. Also available at www.falw.vu.nl/nl/onderzoek/earth-sciences/geo-environmental-science-and-hydrology/hydrology-dissertations/index.asp.

Holwerda, F., R. Burkard, W.E. Eugster, *et al*. (2006). Estimating fog deposition at a Puerto Rican elfin cloud forest site: comparison of the water budget and eddy covariance methods. *Hydrological Processes* **20**: 2669–2692.

Huwe, B., B. Zimmermann, J. Zeilinger, M. Quizhpe, and H. Elsenbeer (2008). Gradients and patterns of soil physical parameters at local, field and catchment scales. In *Gradients in a Tropical Mountain Ecosystem of Ecuador*, eds. E. Beck, J. Bendix, I. Kottke, F. Makeschin, and R. Mosandl, pp. 391–402. Berlin: Springer-Verlag.

Ito, T., S. Shoji, and M. Saigusa (1991). Classification of volcanic ash soils from Konsen district, Hokkaido, according to the last *Keys to Soil Taxonomy* (1990). *Japanese Journal of Soil Science and Plant Nutrition* **62**: 237–247 (in Japanese, with English abstract).

Juvik, J.O., and D. Nullet (1995). Relationships between rainfall, cloud water interception and canopy throughfall in a Hawaiian montane forest. In *Tropical Montane Cloud Forests*, eds. L.S. Hamilton, J.O. Juvik, and F.N. Scatena, pp. 165–182. New York: Springer-Verlag.

Kappelle, M., and A.D. Brown (2001). *Bosques nublados del neotropico*. Heredia, Costa Rica: INBIO.

Klute, A., and C. Dirksen (1986). Hydraulic conductivity and diffusivity: laboratory methods. In *Methods of Soil Analysis,* Part 1, *Physical and Mineralogical Methods*, ed. A. Klute, pp. 687–734. Madison, WI: American Society of Agronomy.

Ko′′hler, L., C. Tobón, K.F.A. Frumau, and L.A. Bruijnzeel (2007). Biomass and water storage dynamics of epiphytes in old-growth and secondary montane cloud forest stands in Costa Rica. *Plant Ecology* **193**: 171–184.

Kubota, T. (1976). Surface chemical properties of volcanic ash soils: especially on phenomenon and mechanism of irreversible aggregation of the soil by drying. *Bulletin of National Agricultural Science* **28**B: 1–74 (in Japanese, with English abstract).

Lal, R. (1987). *Tropical Ecology and Physical Edaphology*. New York: John Wiley.

Lazarovitch, N., A. Ben-Gal, J. Simunek, and U. Shani (2007). Uniqueness of soil hydraulic parameters determined by a combined Wooding inverse approach. *Soil Science Society of America Journal* **71**: 860–865.

Luxmoore, R.J. (1981). Micro-, meso-, and macroporosity of soil. *Soil Science Society of America Journal* **45**: 671–672.

McJannet, D.L., P.G. Fitch, M.G Disher, and J. Wallace (2007a). Measurements of transpiration in four tropical rainforest types of north Queensland, Australia. *Hydrological Processes* **21**: 3549–3564.

McJannet, D.L., J.S. Wallace, and P. Reddell (2007b). Precipitation interception in Australian tropical rainforests. II. Altitudinal gradient of cloud interception, stemflow, throughfall and interception. *Hydrological Processes* **21**: 1703–1718.

Motzer, T., Munz, N., Küppers, M., Schmitt, D., and Anhuf, D (2005). Stomatal conductance, transpiration and sap flow of tropical montane rain forest trees in the southern Ecuadorian Andes. *Tree Physiology* **25**: 1283–1293.

Parasuraman, K., A. Elshorbagy, and S. Cheng (2006). Estimating saturated hydraulic conductivity in spatially variable fields using neural network ensembles. *Soil Science Society of America Journal* **70**: 1851–1859.

Patel, M.S., and N.T. Singh (1981). Changes in bulk density and water intake rate of a coarse textured soil in relation to different levels of compaction. *Journal of the Indian Society of Soil Science* **29**: 110–112.

Ping, C.T., S. Shoji, and T. Ito (1988). Properties and classification of three volcanic ash-derived pedons from Aleutian Islands and Alaska Peninsula, Alaska. *Soil Science Society of America Journal* **52**: 455–462.

Podwojewski, P., J.L. Janeau, and Y. Leroux (2008). Effects of agricultural practices on the hydrodynamics of a deep tilled hardened volcanic ash–soil (Cangahua) in Ecuador. *Catena* **72**: 179–190.

Regalado, C., and A. Ritter (2006). Characterizing water dependent soil repellency with minimal parameter requirement. *Soil Science Society of America Journal* **69**: 1955–1966.

Reynolds, W.D., and D.E. Elrick (1985). In situ measurements of field-saturated hydraulic conductivity, sorptivity, and the alpha-parameter using the Guelph Permeameter. *Soil Science* **140**: 292–302.

Reynolds, W.D., and D.E. Elrick (1991). Determination of hydraulic conductivity using a tension infiltrometer. *Soil Science Society of America Journal* **55**: 633–639.

Reynolds, W.D., D.E. Elrick, and B.E. Clothier (1985). The constant-head well permeameter: effect of unsaturated flow. *Soil Science* **139**: 172–180.

Reynolds, W.D., B.T. Bowman, R.R. Brunke, C.F. Drury, and C.S. Tan (2000). Comparison of tension infiltrometer, pressure infiltrometer, and soil core estimates of saturated hydraulic conductivity. *Soil Science Society of America Journal* **64**: 478–484.

Riverside Salinity Laboratory (2001). *Riverside Salinity Laboratory: Agricultural Research Service*. Available at www.ars.usda.gov/aboutus/aboutus.htm.

Saigusa, M., S. Shoji, and H. Nakaminami (1987). Measurement of water retention at 15 bar tension by pressure membrane method and available moisture of Andosols. *Japanese Journal of Soil Science and Plant Nutrition* **58**: 374–377 (in Japanese, with English summary).

Saghafian, B., P.Y. Julien, and F.L. Ogden (1995). Similarity in catchment response. I. Stationary rainstorms. *Water Resources Research* **31**: 1533–1541.

Santiago, L.S., G. Goldstein, F.C. Meinzer, J.H. Fownes, and D. Mueller-Dombois (2000). Transpiration and forest structure in relation to soil waterlogging in a Hawaiian montane cloud forest. *Tree Physiology* **20**: 673–681.

Schwartz, R.C., P.W. Unger, and S.R. Evett (2000). Land use effects on soil hydraulic properties. In *Proceedings of the 15th Conference of the International Soil Tillage Research Organization*. Washington, DC: Conservation and Production Research Laboratory, U.S. Department of Agriculture.

Schwärzel, K., and P. Punzel (2007). Hood infiltrometer: a new type of tension infiltrometer. *Soil Science Society of America Journal* **71**: 1438–1447.

Shoji, S., M. Nanzyo, and R.A. Dahlgren (1993). *Volcanic Ash Soils: Genesis, Properties and Utilization*. Amsterdam, the Netherlands: Elsevier.

Soil Survey Staff (1996). Keys to Soil Taxonomy, 7th edn. Washington, DC: Natural Resource Conservation Service of the U.S. Department of Agriculture.

Stakman, W.P., and G.G. van der Harst (1969). *Determination of Soil Moisture Retention Curves, Vol. 2, Pressure Membrane Apparatus, Range* pF 3.0 to 4.2. Wageningen, the Netherlands: Institute for Land and Water Management Research.

Stephens, D.B., S. Tyler, K. Lambert, and S. Yates (1983). Field experiments to determine saturated hydraulic conductivity in the vadose zone. In *Role of the Unsaturated Zone in Radioactive and Hazardous Waste Disposal*, ed. J.W. Mercer, pp. 113–126. Ann Arbor, MI: Ann Arbor Science.

Stolte, J., J.I. Freijer, W. Bouten, *et al*. (1994). Comparison of six methods to determine unsaturated soil hydraulic conductivity. *Soil Science Society of America Journal* **58**: 1596–1603.

Tobón, C. (1999). *Monitoring and Modelling Hydrological Fluxes in Support of Nutrient Cycling Studies in Amazonian Rain Forest Ecosystems*. Wageningen, the Netherlands: Tropenbos Foundation.

Tomasella, J., M.G. Hodnett, and L. Rossato (2000). Pedotransfer functions for the estimation of soil water retention in Brazilian soils. *Soil Science Society of America Journal* **64**: 327–338.

Van Genuchten, M.Th. (1980). A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. *Soil Science Society of America Journal* **44**: 892–898.

Vargas, R. (2001). Geología, hidrogeoquímia y modelo conceptual de reservorio para la prefactibilidad del campo geotérmico poco sol, San Ramón – San Carlos, Costa Rica. M.Sc. thesis, University of Costa Rica, San José, Costa Rica.

Vogeler, I., B.E. Clothier, S.R. Green, D.R. Scotter, and R.W. Tillman (1996). Characterizing water and solute movement by time domain reflectometry and disk permeameter. *Soil Science Society of America Journal* **60**: 5–12.

Wang, D., S.R. Yates, and F.F. Ernst (1998). Determining soil hydraulic properties using tension infiltrometers, Time Domain Reflectometry, and tensiometers. *Soil Science Society of America Journal* **62**: 318–325.

Warrick, A.W., and D.R. Nielsen (1980). Spatial variability of soil physical properties in the field. In *Applications of Soil Physics*, ed. D. Hillel, pp. 319–344. New York: Academic Press.

Wooding, R.A. (1968). Steady infiltration from a shallow circular pond. *Water Resources Research* **4**: 1259–1273.

Woolhiser, D.A., R.E. Smith, and J.V. Giraldez (1996). Effects of spatial variability of saturated hydraulic conductivity on Hortonian overland flow. *Water Resources Research* **32**: 671–678.

Xiang, J. (1994). Improvements in evaluating constant-head permeameter test data. *Journal of Hydrology* **162**: 77–97.

Xiang, J., B.R. Scanlon, W.F. Mullican, L. Chen, and R.S. Goldsmith (1997). A multistep constant-head borehole test to determine field saturated hydraulic conductivity of layered soils. *Advances in Water Resources* **20**: 45–57.

Zadroga, F. (1981). The hydrological importance of a montane cloud forest area of Costa Rica. In *Tropical Agricultural Hydrology*, eds. R. Lal and E.W. Russell, pp. 59–73. New York: John Wiley.

Ziegler, A.D., T.W. Giambelluca, L.T. Tran, *et al*., (2004). Hydrological consequences of landscape fragmentation in mountainous northern Vietnam: evidence of accelerated overland flow generation. *Journal of Hydrology* **287**: 124–146.

Zimmermann, B., and H. Elsenbeer (2008). Spatial and temporal variability of soil saturated hydraulic conductivity in gradients of disturbance. *Journal of Hydrology* **361**: 78–95.