

6 Tropical Forest Hydrology

T. Kumagai^{1*}, H. Kanamori¹ and N.A. Chappell²

¹Nagoya University, Nagoya, Japan; ²Lancaster University, Lancaster, UK

6.1 Introduction

One may imagine that tropical forest regions are in general characterized by higher incident radiative energy, constantly higher temperature and a large amount of ecosystem water resources, enough for supporting their higher primary productivity and active water cycling. In reality, tropical forests play a significant role in the global carbon budget (e.g. Beer *et al.*, 2010; Pan *et al.*, 2011) and are a major source of global hydrological fluxes, profoundly influencing both global and regional climates (e.g. Avissar and Werth, 2005; Spracklen *et al.*, 2012; Poveda *et al.*, 2014).

Tropical forest regions may be roughly classified into two types based upon seasonal variations in precipitation, although each region can be classified into more climate and forest types (e.g. Walsh, 1996; Tanaka *et al.*, 2008): (i) tropical evergreen rainforest with a 'rainforest climate'; and (ii) tropical seasonal forest with a 'monsoon climate' (Kumagai *et al.*, 2009). Tropical evergreen rainforests exist only in regions of ample water resources (Kumagai *et al.*, 2005; Kumagai and Porporato, 2012b), while tropical seasonal forests have to cope with seasonal droughts (e.g. Vourlitis *et al.*, 2008; Miyazawa *et al.*, 2014; Kumagai *et al.*, 2015). Thus, since

the existence of these tropical forests achieves a delicate balance in terms of their ecosystem water resources, hydrological change in either region could result in significant impacts on ecological patterns and processes (Malhi *et al.*, 2009; Phillips *et al.*, 2009; Kumagai and Porporato, 2012a), in turn affecting feedbacks to the atmosphere (Meir *et al.*, 2006; Bonan, 2008; Kumagai *et al.*, 2013; van der Ent *et al.*, 2014). Here, we should note that recent evidence shows marked change in regional climate will occur first in the tropics, making tropical forest ecosystems particularly vulnerable in the future (Mora *et al.*, 2013).

Humans have been modifying the tropical forest land cover for food and energy production and for the development of the tropical countries. Consequently, such modifications (i.e. land-use changes) are being combined with climate change and should be anticipated to impact the regional hydro-climate as well as the local freshwater resources (Bruijnzeel, 1990; Giambelluca, 2002; van der Ent *et al.*, 2010; Wohl *et al.*, 2012; Kumagai *et al.*, 2013). What distinguishes the current modifications are the intensity and global reach, where the entire hydro-climate is now subject to these modifications. The consequences of land-use and climate change in the tropical forest regions are considered among the

*Corresponding author; e-mail: toomoomikumagai@gmail.com

greatest environmental concerns for the survival of the human population today, and include changes in streamflow and flood frequency, loss of productive soils, changes in nutrient fluxes, a warmer and drier climate, and concomitant changes in hydrological ecosystem services at the local scale (e.g. Bruijnzeel, 2004; Chappell *et al.*, 2004; Krusche *et al.*, 2011; Lawrence and Vandecar, 2015). To tackle all aspects of these hydrological problems, we have broad recognition and tacit acceptance that progress on these complex issues benefits from fundamental knowledge on all hydrological components in various types of tropical forest and synthesizing them in the global implications (see Bruijnzeel, 2004).

The main objective of this chapter, therefore, is to provide the state of knowledge on the characteristics of the hydrological components of precipitation, evapotranspiration and streamflow generation in tropical forest regions by a review of cornerstone tropical literature and a pan-tropical perspective partly via the use of pan-tropical maps of the hydrological components. Then, we suggest research needs and strategies for the study of tropical forest hydrology in the context of a dramatically changing world.

6.2 Pan-Tropical Climatic Regime and Forest Type

Pan-tropical maps of precipitation characteristics are derived from the PREC/L long-term gridded precipitation data set (Chen *et al.*, 2002) (Plate 5). High radiative energy received around the equator generates the updraft of hot and humid air. The convergence of the hot and humid air (i.e. the low atmospheric pressure system) results in the belt of precipitation (Plate 5a) associated with the Inter-Tropical Convergence Zone (ITCZ) around the equator. Updraft in the ITCZ moves both northward and southward in the tropopause, and descends around the Tropic of Cancer and the Tropic of Capricorn. Around both tropics (the so-called 'horse latitudes') the divergence of hot and dry air typically creates arid areas and deserts (Plate 5a). This circulation of air masses ascending around the equator and descending around tropics is called the Hadley circulation.

According to the Köppen climate classification, tropical climates are defined as the condition in

which air temperatures do not fall below 18°C in the coolest month where the annual mean precipitation is more than the Köppen arid boundary ($P_k = 20 \times (T_y + x)$ in mm, where T_y is annual mean air temperature (°C) and x is an index representing the precipitation pattern such as 0, 7 and 14 for dry summer and wet winter, wet throughout the year, and dry winter and wet summer, respectively). These climates are further subdivided into rainforest (Af), monsoon (Am) and savannah (Aw) classes. The rainforest climate occurs in the zone where all 12 months have mean precipitation of at least 60 mm. The monsoon climate and the savannah climate have mean precipitation less than 60 mm in the least precipitation month, but they have the least monthly precipitation more and less than $100 - 0.04 \times$ (annual mean precipitation in mm), respectively.

As Feng *et al.* (2013) suggested, the areas with the largest amount of precipitation always exist in the regions with the most aseasonal precipitation patterns, but not vice versa (Plate 5). The rainforest climate (Af) is characterized by both a large annual amount of precipitation (Plate 5a) and little or negligible seasonal variation in precipitation (Plate 5b), caused by a stationary low atmospheric pressure system around the equator. Tropical evergreen forests can exist in this region because of their constant water requirements throughout the year. High rainfall intensity of short duration is distinctive in the rainforest climate, and in both diurnal and seasonal variation senses, there is strongly spatial heterogeneous rainfall distribution over tropical rainforests at the scale of single to thousands of square kilometres (Krusche *et al.*, 2011; Kumagai and Kume, 2012; Kanamori *et al.*, 2013).

The monsoon (Am) and the savanna (Aw) climates are, by contrast, characterized by their degree of seasonality in precipitation, caused primarily by seasonal fluctuations of the ITCZ. It should be noted that annual precipitation decreases with the increased seasonality of precipitation, but not vice versa (see Feng *et al.*, 2013; Plate 5). There is a less intense dry season and a larger amount of precipitation in the monsoon climate than in the savannah climate. Tropical seasonal forests can exist in areas of the monsoon with climate adaptations that cope with water stress in the dry season, e.g. full- or semi-deciduous traits, seasonal variations in

stomatal behaviour, etc. This, in turn, means that tropical seasonal forest can vary from evergreen to deciduous forests corresponding to the local precipitation regimes such as influenced by the altitude and topography (e.g. Tanaka *et al.*, 2008).

The tropics are alternatively divided into three regions based on biotic and geological history (US DOE, 2012): (i) the neo-tropical ecozone (NEO) of South America, Central America and the Caribbean; (ii) the Afro-tropical ecozone (AFR) of sub-Saharan Africa; and (iii) the Indo-Malay-Australasia tropical ecozone (IMA) including regions of India, South-East Asia, southern China, New Guinea and northern Australia. The forests of the Amazon Basin in the neo-tropics represent the single largest block of intact tropical forests, occupying ~40% of forest biomass in the tropics (Saatchi *et al.*, 2011). South and Central American and Caribbean forests often experience hurricane-induced disturbance (e.g. Boose *et al.*, 1994, 2004; Negrón-Juárez *et al.*, 2010). Among the Afro-tropical forests, those of the Congo Basin represent the second-largest intact block of tropical forests (Pan *et al.*, 2011). Central African forests do not experience tropical cyclones but are still subject to large storm events, and suffer from severe drought and special wet periods particularly during the El Niño and the La Niña phenomena (US DOE, 2012). While the dominant plant family in neo-tropical and Afro-tropical forests is *Leguminosae*, within South-East Asian forests *Dipterocarpaceae* dominate. Dipterocarps are among the most valuable timber species globally, leading to intense logging pressure within the Indo-Malay-Australasia tropical forests. Indeed over the past 50 years, more timber was exported from Borneo Island in this region than from neo-tropical and Afro-tropical regions combined (Curran *et al.*, 2004). Although South-East Asian tropical forests represent only about 11% of the world's tropical forests in terms of area, they have the highest relative deforestation rate in the tropics (e.g. Canadell *et al.*, 2007; Pan *et al.*, 2011). An analysis of climatic trends in global tropical rainforest regions over the period 1960–1998 showed that precipitation has declined more significantly in South-East Asia than in Amazonia, and that the El Niño–Southern Oscillation (ENSO) is a particularly important driver of drought in South-East Asia (Malhi and Wright, 2004).

6.3 Pan-Tropical Evapotranspiration

Evapotranspiration, i.e. water vapour evolution from the forest canopy, can be simply partitioned into wet canopy evaporation or canopy interception and dry canopy evaporation or transpiration. Kume *et al.* (2011) reviewed tropical forest canopy interception ratios, i.e. the wet canopy evaporation divided by gross precipitation, from the data of 40 tropical forests including both rainforests and seasonal forests, and reported that most ratios measured in tropical forests were in the range of 10–20% with a mean of 17%. On the other hand, the reported transpiration rates in tropical forests have a large temporal and spatial variation in value as 2.3–4.6 mm/day (e.g. Shuttleworth *et al.*, 1984; Roberts *et al.*, 1993; Cienciala *et al.*, 2000; Kumagai *et al.*, 2004). Bruijnzeel (1990) suggested that in humid tropical forests, the average annual transpiration was 1045 mm (range 885–1285 mm). The resultant variation in annual tropical forests' evapotranspiration ranges from about 1000 to 1500 mm and the ratio of annual evapotranspiration to annual precipitation ranges mostly between 30 and 90% (Kume *et al.*, 2011). It is worth noting that Kume *et al.* (2011) found that apparently in tropical forests when the annual precipitation (P) is <2000 mm, the evapotranspiration increases with increases in P , and when $P > 2000$ mm, the evapotranspiration reaches the plateau of about 1500 mm.

There are more complex situations controlling tropical forests' evapotranspiration. Trees in the tropical monsoon (Am) climatic regions tend to cope with seasonal drought by stomatal control, leaf-fall, root water uptake at deeper soil depths and so forth (e.g. Igarashi *et al.*, 2015). Many pristine tropical forests have been cleared for conversion to plantations such as rubber trees and oil palms (e.g. Carlson *et al.*, 2012; Fox *et al.*, 2012). Such plantations' water use is comparable to or more than that of the original vegetations (e.g. Guardiola-Claramonte *et al.*, 2010; Kumagai *et al.*, 2015). Here, we should note that significant portions of the cleared tropical land revert quickly to secondary vegetation and, in terms of hydrological characteristics, increasingly resemble original forest with time (Giambelluca, 2002). Giambelluca *et al.* (2003) pointed out that evapotranspiration from fragmented forest tends to be enhanced by conditions in surrounding

clearings. Tropical montane cloud forests, which are fog- and cloud-affected, need to be paid special attention on their hydrometeorology (Bruijnzeel *et al.*, 2011): small and large amounts of transpiration and canopy interception due to the small water vapour deficit and the cloud-water, respectively.

Despite the complexity in generalizing mechanisms of evapotranspiration, we attempt comparison of the tropical forests with other regions' evapotranspiration: annual evapotranspiration (E) from ground observations (e.g. watershed water balance, micrometeorological measurements and soil-water balance) were classified by annual mean temperature (T) and represented as a function of latitude (Fig. 6.1, adapted from Komatsu *et al.*, 2012). Despite a large variation in E at each latitude (~ 500 mm), which might be caused by altitude, forest type and local climate of the forest sites, differences in hydrological observation methods and so forth, there is a strong linear relationship between latitude and E . In humid climates of the ITCZ, the high P is expected to cause higher E (Fig. 6.1). Note that there is a large variation in E in the tropical forest regions, but many can be explained by the

difference in the T , probably because of the altitude effects.

From this relationship between annual P and T , Komatsu *et al.* (2012) modified the Zhang *et al.* (2001) evapotranspiration model where:

$$E = P \frac{1 + w(E_0 / P)}{1 + w(E_0 / P) + (P / E_0)} \quad (6.1)$$

in which

$$E_0 = 0.488T^2 + 27.5T + 412 \quad (6.2)$$

where w is a coefficient representing plant water availability ($= 2.0$). E_0 is the potential evaporation, which was defined as a constant value calculated by Priestley and Taylor's (1972) equation in Zhang *et al.* (2001) but was modified in Komatsu *et al.* (2012) so that it can appropriately describe the temperature effect. Eqns 6.1 and 6.2 successfully reproduce E as a function of latitude (Fig. 6.1). In the tropical rainforest and monsoon climates it suggests that higher E eventuates through the conditions of a plentiful P and higher T .

Theoretical (Eqns 6.1 and 6.2) and observed relationships between annual P and E are

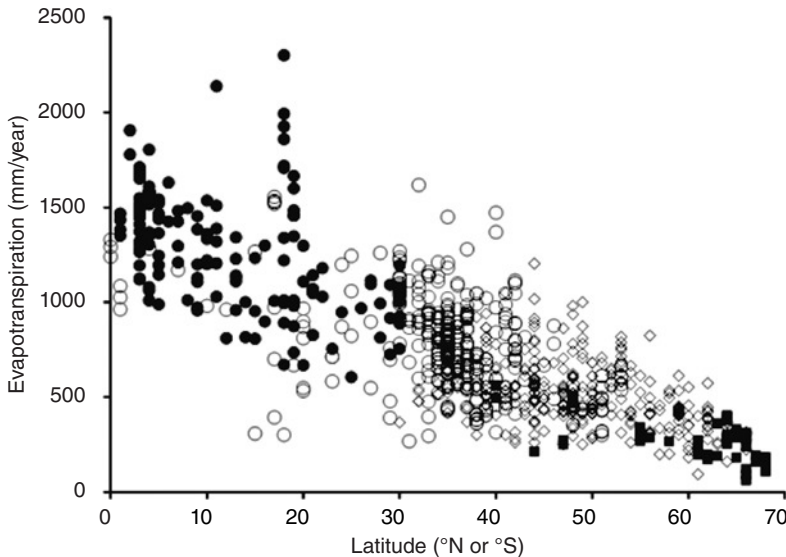


Fig. 6.1. Annual evapotranspiration as a function of latitude, classified by annual mean temperature (T ; °C): closed circles = $T > 20$; open circles = $10 < T < 20$; open diamonds = $0 < T < 10$; closed squares = $T < 0$. (Data on evapotranspiration, latitude and temperature taken from Komatsu *et al.*, 2012.)

investigated in Fig. 6.2a1–c1. The shallower slope in the relationship between E and P in the neo-tropics (Fig. 6.2a1) and the Indo-Malay-Australasia tropics (Fig. 6.2c1) compared with the steeper relationship in the Afro-tropics (Fig. 6.2b1) can be explained by the saturation curve, Eqn 6.1. It is also surprising that such a simple formulation as Eqn 6.1 has an ability to describe the pan-tropical E characteristics because Fig. 6.2 contains data not only from natural seasonal and rain forests but also secondary forests and plantations. Further, it is interesting to note that from an interpretation of Budyko's vegetation categorization using the relationships among the radiation dryness index, net radiation

and vegetation types (see Budyko and Miller, 1974), tropical forests can exist within the ranges of annual mean net radiation 110–133 W/m^2 and annual mean precipitation 1370–5000 mm. The lowest precipitation limit for the presence of tropical forest (i.e. 1370 mm/year) is found to be broadly comparable with the lowest value of P in each ecozone (Fig. 6.2a1–c1).

The local water-use ratio (LEUR) is defined as E divided by P , and represents the ratio of P water recycling from E . Figure 6.2a2–c2 shows these data for each tropical ecozone. The LEUR data show a significant decrease with P for all ecozones. This implies that in areas with smaller P (mainly in AFR), the P is effectively recycled

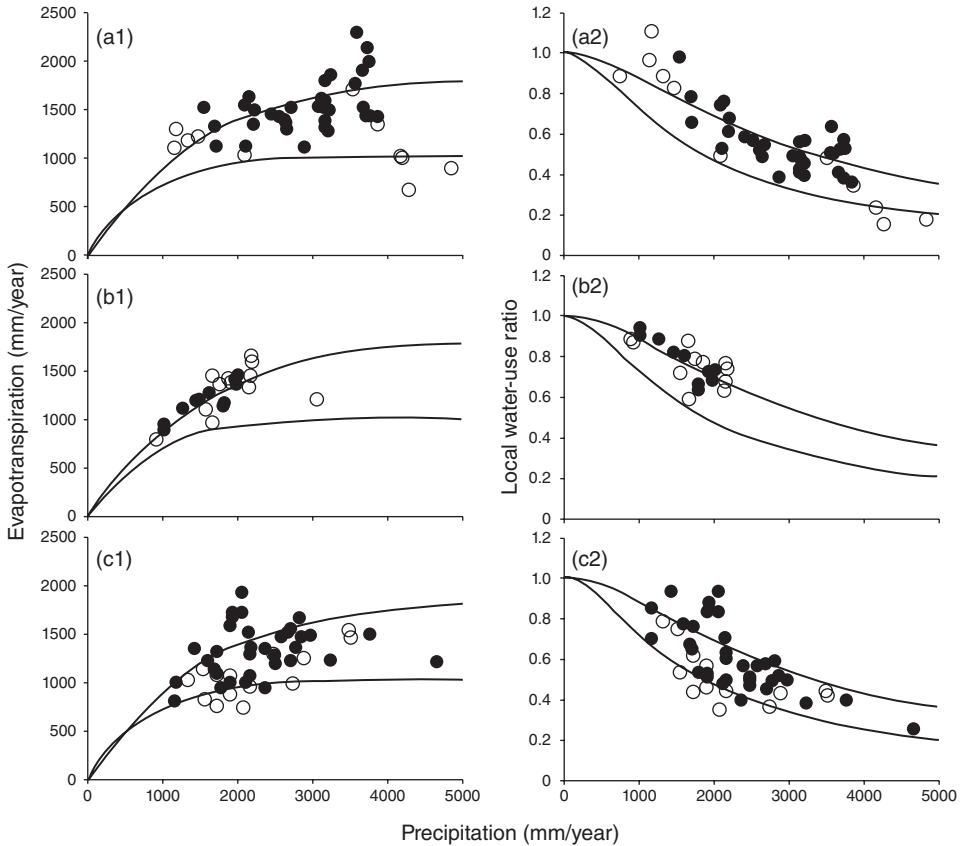


Fig. 6.2. Relationships between precipitation and annual evapotranspiration (a1, b1, c1) and local water-use ratio (a2, b2, c2) defined as evapotranspiration divided by precipitation, classified by annual mean temperature (T ; $^\circ\text{C}$): closed circles = $T > 25$; open circles = $20 < T < 25$. Solid lines denote the theoretical relationships: upper and bottom lines for each panel assume $T = 30$ and 15 , respectively. (a1, a2) neo-tropical ecozone; (b1, b2) Afro-tropical ecozone; and (c1, c2) Indo-Malay-Australasia tropical ecozone. (Data on evapotranspiration and precipitation taken from Komatsu *et al.*, 2012.)

from the E so that the vegetation is regulating its own regional water resources. In areas with larger P (such as NEO and IMA), P is also supplied by moist air from the surrounding terrain and oceans.

The atmosphere–land surface water balance method (see Oki *et al.*, 1995) with global data on P (Plate 5) and atmospheric water vapour divergence derived from ERA-Interim gridded four-dimensional meteorology data set (Dee *et al.*, 2011) produced a pan-tropical map of E and LEUR. Although E is well reproduced in areas with a relatively small amount of P , it is likely that large P in the method has induced an overestimation of E in such areas. Nevertheless, characteristics of spatial variation in E are well represented. Comparatively small E can be seen in the Afro-tropics, and a much larger E in the islands of Borneo and New Guinea in the Indo-Malay-Australasian tropics and in Amazonian headwaters in the neo-tropics. Areas with lower LEUR values tend to overlap areas with larger P (compare with Plate 5a) and this is consistent with the patterns shown in Fig. 6.2a2–c2. The exception is Borneo Island (IMA ecozone) where despite a plentiful P , higher values of LEUR are estimated due to near-zero annual atmospheric water vapour divergence/convergence (Kumagai *et al.*, 2013). Here, we should note that for considering the extent to which P relies on terrestrial E (i.e. moisture recycling), the role of global wind patterns, topography and land cover should be interpreted more in the context of continental moisture recycling (van der Ent *et al.*, 2010). Notably, Poveda *et al.* (2014) examined how ‘aerial river’ pathways modified by the effects of topography, orography and land cover types contribute to precipitation patterns in tropical South America.

6.4 Pan-Tropical Streamflow Generation

Figure 6.3 shows a plot of annual streamflow per unit drainage area (the strict definition of term ‘runoff’) against latitude (determined at the confluence with the ocean; adapted from Wohl *et al.*, 2012). The streamflow for December–February (winter in the northern hemisphere) and June–August (winter in the southern

hemisphere) averaged per latitude was also estimated using the atmosphere–land surface water balance method with assumptions that deep groundwater flow all contributes to streamflow at the scale of large watersheds and changes in calculation-domain water storage can be neglected (Oki *et al.*, 1995) and added to Fig. 6.3. The negative values of computed 3-month streamflow are probably attributable to the role of dynamic subsurface storages. It is, however, apparent that the annual streamflow from tropical basins is typically much greater than that from temperate basins due primarily to the greater precipitation amounts in the humid tropics. Further, the range in annual observed streamflow is much larger within the latitudes 0 to 23.4° due to the presence of extensive areas with dry climates (BW and BS) in the tropics.

The greater streamflow present within the parts of the tropics with a rainforest climate (Af) and monsoon climate (Am) indicates that considerably more precipitation travels through watersheds towards streams than at other latitudes. The greater water flows within such tropical basins mean that the magnitudes of chemical and particle transport are likely to be greater and the watershed systems more sensitive to disturbance (Wohl *et al.*, 2012). Consequently, tropical forest hydrology has implications for the other scientific disciplines of biogeochemistry and geomorphology.

Some 89% of the channel network of the globe’s streams/rivers comprises first-, second- and third-order channels (Table 2 in Downing *et al.*, 2012). This means the most streamflow is generated in the network of such low-order channels. Experimental watersheds typically comprise channels of first- to third-order size, and so are ideal locations for the study of the pathways of rainwater to the channel network via surface and/or subsurface pathways. These routes of water migration to channels are known as the pathways of ‘streamflow generation’ or simply ‘runoff pathways’ (Bonell, 2004; Burt and McDonnell, 2015). Most experimental watersheds used for the study of runoff pathways are located in temperate regions, with very few in tropical regions (Bonell, 2004; Burt and McDonnell, 2015). Figure 6.4a shows the locations of some key experimental watersheds with a long history of research on streamflow generation pathways that are located in the

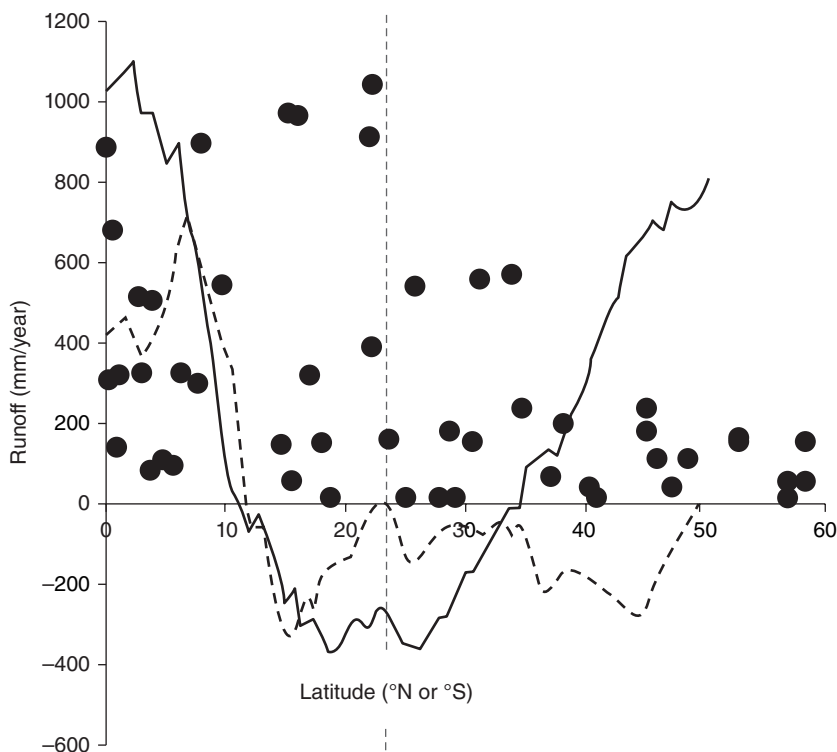


Fig. 6.3. Annual-scaled streamflow per unit drainage area as a function of latitude at the river mouth (circles) adapted from Wohl *et al.* (2012). Relationships between latitude and annually scaled and latitude-averaged streamflow computed from the atmosphere–land surface water balance method for December–February in the northern hemisphere (solid line) and June–August in the southern hemisphere (broken line) are also shown. Vertical dotted line denotes the latitude of the tropics north and south of the equator (23.4°).

tropics. Most of these watersheds are located beneath tropical forests, so the findings are most pertinent to tropical forest environments.

Streamflow generation pathways that have been observed within these tropical forest environments include:

- 1. Infiltration-excess overland flow.** This water flow on slopes outside channels is caused by precipitation falling at a rate faster than the local coefficient of permeability at the ground surface (equivalent to the ‘infiltration capacity’ or surface ‘saturated hydraulic conductivity’).

- 2. Saturation overland flow by direct precipitation.** Where rainfall falls on to ground at a rate less than that of the infiltration capacity, but where pores are already saturated, then no further infiltration can take place and new water travels over the surface.

- 3. Subsurface flow.** Where rainfall infiltrates, some will evaporate from the soil or support transpiration from vegetation; the remainder will travel towards streams below the surface. Most of this water will enter the streams via the channel bed and banks, but some will return to the surface prior to reaching a channel (so-called *return flow*) and flow over the surface as *saturation overland flow*. Flow beneath the surface may be very shallow where lithomorphic soils overlie an impermeable geology, but may be over 100 m deep where permeable soils overlie permeable geology (whether unconsolidated materials or rock).

Some ambiguity in the definition of water pathways arises from the definition of the ground surface. Some scientists define overland flow as

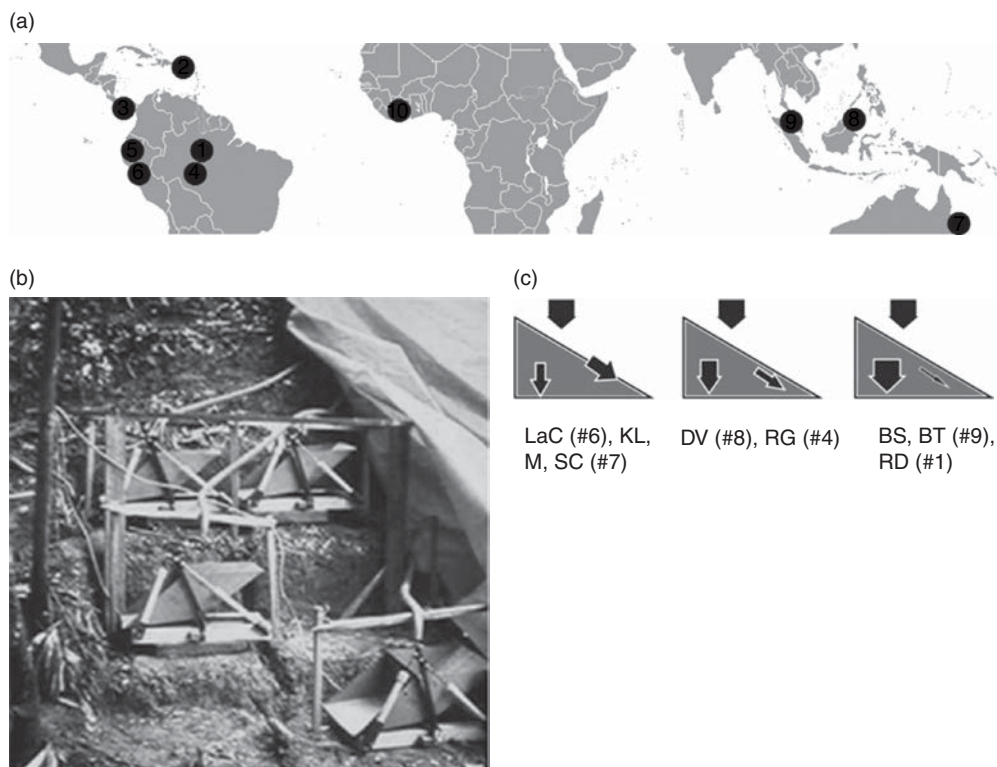


Fig. 6.4. (a) Location of ten experimental basins in the tropics with a long history of research into streamflow generation pathways: #1 = Reserva Ducke (Brazil); #2 = Bisley II (Puerto Rico); #3 = Barro Colorado & Lutzito (Panama); #4 = Rancho Grande & Jurena Ultisol (Brazil); #5 = Rio San Francisco (Ecuador); #6 = La Cuenca (Peru); #7 = South Creek (Australia); #8 = Danum Valley (Malaysia); #9 = Bukit Tarek & Bukit Berembun (Malaysia); #10 = Tai Forest (Ivory Coast) (for an extensive list of publications for these and other experimental sites in the tropics, see *Elsenbeer, 2001; Bonell, 2004; Hugenschmidt et al., 2014; Barthold and Woods, 2015*). (b) The 'throughflow trough' system used by *Bonell and Gilmour (1978)* to demonstrate the presence of shallow lateral flows in the organic and mineral soils at the South Creek Experimental Watershed (#7), Queensland, Australia. (c) Inferred dominant streamflow generation pathways presented in *Elsenbeer (2001)*. Experimental watershed LaC = La Cuenca (#6); KL = Kiani Lestari; M = Mendolong; SC = South Creek (#7); DV = Danum Valley (#8); RG = Rancho Grande (#4); BS = Bukit Soeharto; BT = Bukit Tarek (#9); RD = Reserva Ducke (#1) (see *Elsenbeer, 2001* for details).

water moving over the surface of a mineral A (or E) soil horizon, while others use the definition of water moving laterally above the typically overlying organic horizons (i.e. L, H and/or O horizons).

Streamflow generation studies in the tropics that have directly observed the presence of overland flow where the rainfall intensity exceeds the infiltration capacity are very limited, but the Tai Forest basin is a good example (*Bonell, 2004*). Studies that have observed overland flow on permeable but saturated soils include the

South Creek (*Bonell and Gilmour, 1978*) and La Cuenca basins (*Elsenbeer and Vertessy, 2000*). Several studies have direct evidence of lateral (downslope) flows within the subsurface generated during rainstorms. *Bonell and Gilmour (1978)* observed the presence of these flows using so-called 'throughflow troughs' (*Fig. 6.4b*), while *Chappell and Sherlock (2005)* observed the presence of these flows by monitoring tracer migration.

The more pertinent question is not whether a particular pathway is present or not, but

whether it is the *dominant pathway* producing >50% of the observed total streamflow, particularly during storm events (i.e. without any manipulation of hydrographs by separation methods). Too often researchers infer the dominance of a particular pathway simply from the observed presence of the pathway, rather than relating measured *flows per unit basin area from that pathway* with those observed in the *stream per unit basin area*. For example, the measured lateral flows per unit basin area of Gilmour *et al.* (1980) are only a tiny fraction of the observed streamflow per unit basin area. Consequently, the importance of measured near-surface flows at this site (and many other sites) has been overemphasized in comparison to the unmeasured flows within deeper soil and unconsolidated rock strata. The resultant ambiguity and misinterpretation of the dominant pathway is amply illustrated by inconsistencies in the inferred dominant pathways presented in the reviews of Elsenbeer (2001), Bonell (2004) and Barthold and Woods (2015) as shown in Fig. 6.4c and Table 6.1.

A further area of concern is the focus of most studies on pathways in the solum (i.e. A and B soil horizons) alone, as highlighted by Bonell and Balek (1993) and Bonell (2004). There is an increasing awareness that at some tropical sites, soils may be developed on unconsolidated geological materials that are permeable and have deeper pathways within these strata. Examples of experimental watersheds developed on these deeper porous media in the tropics include the Lake Calado microbasin near Reserva Ducke basin in Brazil (Lesack, 1993); the Jungle Falls basin in Singapore (Chappell and Sherlock, 2005; Rahardjo *et al.*, 2010); the Bukit Timah (Noguchi *et al.*, 2005) and Bukit Berembun (Chappell *et al.*, 2004) basins in Malaysia; and the O Thom II basin in Cambodia (Shimizu *et al.*, 2007). Equally, other low-order basins in the tropics may be located on rock aquifers or rocks with fracture systems that produce very deep water pathways. The Arbole-da basin near La Selva, Costa Rica (Genereux *et al.*, 2005) has such pathways. To alert researchers to the potential presence and role of these deeper streamflow generation pathways, Chappell *et al.* (2007) developed a perceptual model of runoff pathways, where a type III system has a dominance of pathways via

Table 6.1. Inferred dominant streamflow generation pathways presented in Table 14.1 of Bonell (2004). See Bonell (2004) for further details of the experimental watersheds and pathway definitions.

Dominant pathway	Experimental watershed ^a
Predominantly vertical pathways	Reserva Ducke (#1), Bukit Tarek (#9), Fazenda Dimona, Mgera
Predominantly lateral pathways (infiltration-excess overland flow)	Tai Forest (#10)
Predominantly lateral pathways (saturation-excess overland flow)	South Creek (#7), Barro Colorado (#3), La Cuenca (#6), Maburac
Predominantly lateral pathways (subsurface stormflow within the soil)	Danum Valley (Sungai Baru Barat) (#8), Rancho Grande (#4), Bisley II (#2), Dodmane, Kannike, Ife, ECEREX
Predominantly lateral pathways (subsurface stormflow at the soil–bedrock interface)	Kuala Belalong

^aEach # represents the location in Fig. 6.4a.

unconsolidated geological materials and a type IV system has a dominance of pathways via rock aquifers or fracture systems (Fig. 6.5). Experimental basins lacking any evidence of major pathways through geological strata (solid rock or unconsolidated materials) are classified as either type II systems (where a B soil horizon is present, e.g. Chappell *et al.*, 1998) or type I systems (where a lithomorphous soil is developed on steep, impermeable mountain slopes). The potential presence of deeper pathways within existing experimental watersheds needs to be a key focus for new research. With more complete observational evidence, including information gained from both hydrometric and tracer studies (as recommended by Barthold and Woods, 2015), researchers may be closer to developing a unified, numerical model of the dominant pathways of streamflow generation applicable across all tropical forest environments.

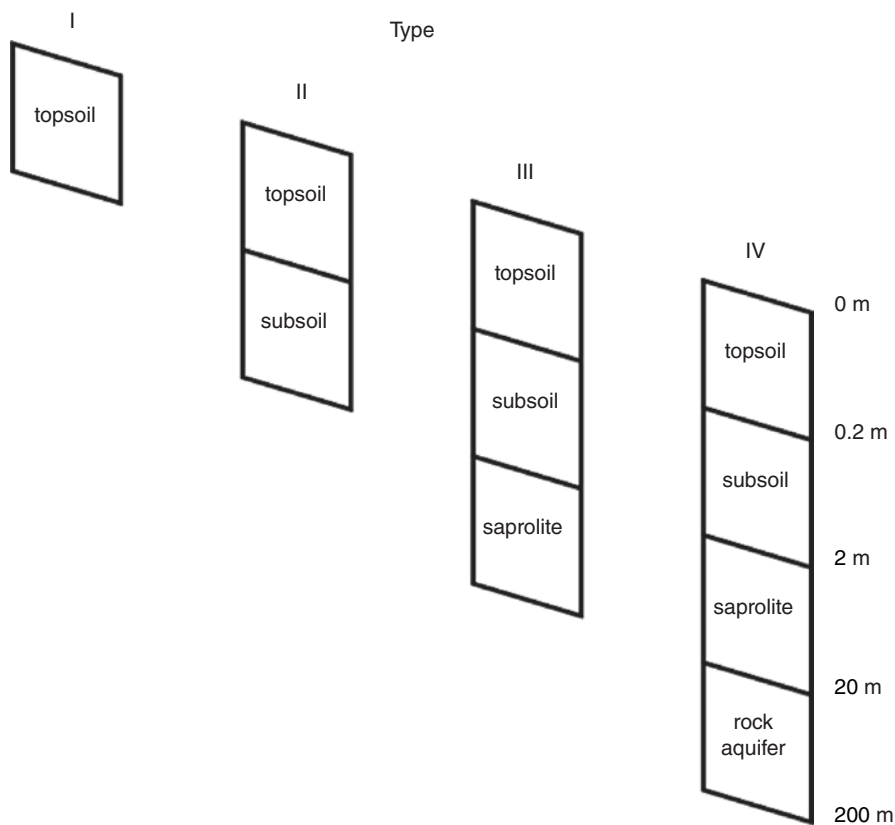


Fig. 6.5. Schematic representation of water pathway systems within tropical forests classified according to the presence of as many as four strata (soil horizon or topsoil (A); soil horizon or subsoil (B); unconsolidated geological materials; and solid rock) with the names type I to IV. Example depths in the logarithmic depth scale are given. (Adapted from Chappell *et al.*, 2007 and US DOE, 2012.)

6.5 Research Needs

The projected growth in atmospheric greenhouse gases within the coming century, as predicted by the Intergovernmental Panel on Climate Change's (IPCC) A1B (Balanced across all sources) scenario, will significantly increase tropical surface temperatures ranging from ~3 to 5°C for South-East Asia, the Amazon and West Africa (IPCC, 2007). Also, the A1B scenario predicts concomitant modifications to precipitation patterns: a general increase in precipitation for West Africa, intensification of seasonality of precipitation for South-East Asia (i.e. more and less precipitation in the wet and dry seasons, respectively) and a general decrease in precipitation for the southern and eastern Amazon. More dramatically, as an instance, the 1997–1998

El Niño was the strongest in the 20th century and its associated drought in Borneo Island in South-East Asia was the most severe (statistically, a drought such as in 1998 may occur once in ~360 years). Tree mortality rates during that drought were 6.37%/year, as compared with 0.89%/year during the pre-drought period (1993–1997) in the studied forest site in western Borneo (Nakagawa *et al.*, 2000). Global warming is likely to cause changes in Pacific regional climate that might alter ENSO activity in the future (e.g. Timmermann *et al.*, 1999). While it is not still clear how ENSO will be affected by global warming, it is possible that the frequency or amplitude of ENSO events could increase (Collins *et al.*, 2010).

In addition, the tropics are known to be a very active domain in terms of changes in land

cover. According to Hansen *et al.*'s (2013) investigation using Earth observation satellite data from 2000 to 2012, the tropics were the only domain to exhibit a statistically significant increasing trend in annual forest loss, 2101 km²/year, and tropical rainforests loss was 32 % of the global forest loss. South American tropical dry forests have been lost at the highest rate in the lost tropical forests. Furthermore, Hansen *et al.* (2013) confirmed that tropical forest loss rate in Brazil has decreased from 40,000 km² in 2003–2004 to 20,000 km² in 2010–2011, while forest loss in Indonesia increased drastically from 8000 km² in 2000–2001 to 20,000 km² in 2010–2011, indicating that an increase in Asian tropical forest loss 'compensates' a decrease in Amazonian tropical forest loss.

In the humid tropics, where the deforestation and land-use change are still ongoing and there is plenty of precipitation, a combination of climate change and drastic changes in land cover must induce pressure for freshwater resources and loss of soil via changes in local hydrological processes. Tropical forest evapotranspiration is generally greater than evapotranspiration from grasslands and thus changing land cover such as from forests to pastures would reduce the evapotranspiration and increase streamflow, resulting, in some cases, in increasing flood frequencies (see Bruijnzeel, 2001, 2004). The generation of localized infiltration-excess overland flow on compacted soil surfaces caused by forestry operations accelerates the geomorphological process of soil erosion (see Bruijnzeel, 2004; Sidle *et al.*, 2006; Sidle and Ziegler, 2012). Such land degradations would become increasingly worse by altered precipitation regimes through climate change (see Peña-Arancibia *et al.*, 2010; Wohl *et al.*, 2012). On the other hand, we should note that the hydrological properties of the secondary vegetation such as evapotranspiration and surface infiltration may quickly resemble those of the original forest again (see Giambelluca, 2002; Bruijnzeel, 2004).

From the global perspective, once more it should be emphasized that the tropical forests are a major source of global hydrological fluxes and thus changes in evapotranspiration rates could significantly impact both global and regional climates, in turn potentially affecting feedbacks to the atmosphere (see Bonan, 2008). Many previous research contributions to the knowledge on hydro-climate in the tropics suggest: forest cutting is

more effective on local and global climate under maritime conditions than continental conditions (van der Molen *et al.*, 2006); rainwater is recycled earlier by wet canopy evaporation than via transpiration (van der Ent *et al.*, 2014); and in the maritime continent, changes in sea surface temperature influence precipitation regimes more than land cover changes (Bruijnzeel, 2004). Further, large-scale tropical deforestation and selective logging could result in warmer and drier conditions not only at the local scale, but also the teleconnections from converted tropical lands could pose a considerable risk to agriculture in other regions, due to impacts on precipitation against a background of warmer temperatures (see Lawrence and Vandecar, 2015).

As the first step for considering the future hydrological impact in the tropics, we need an understanding of the current and basic tropical hydrological cycling – and further, of the hydrological interactions among the earth surface and subsurface, vegetation and atmosphere – based on long-term and networked data acquisition and organization. As Wohl *et al.* (2012) pointed out, field-based hydrological measurements in many tropical countries have been less explored than those in the temperate regions and, to make matters worse, are waning. We should note that the lack of long-term observations homogenized throughout the tropics leads to a failure to validate hydro-climate models and thus the impossibility to extrapolate the future tropical forest hydrology by output of the models, which must be built referring to reliable observations and should use the observations as inputs. Besides making more effort for organizing a field-based hydrological observation network over the tropics, as shown in this chapter, remote sensing technologies and global-scale climate and geo-information databases can serve among the most promising tools to complement the lack of tropical field-observation sites.

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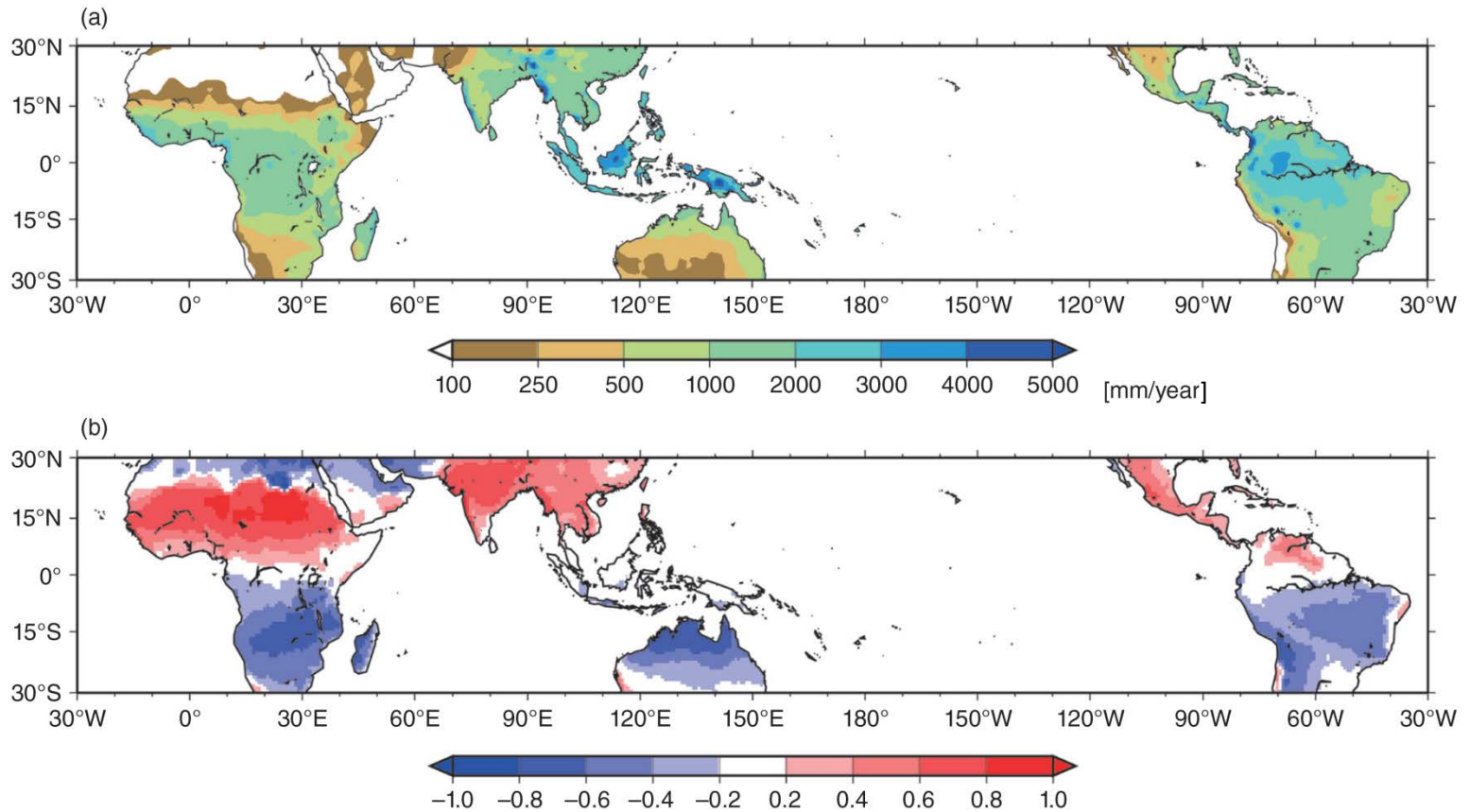


Plate 5. Pan-tropic maps of (a) annual mean precipitation and (b) precipitation seasonality represented by the ratio of difference in precipitation between June–August and December–February to annual precipitation. (Constructed using the National Oceanic and Atmospheric Association’s (NOAA) PRECipitation REConstruction over Land (PREC/L) with values averaged over 1979–2012.)