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Hydrological functions of tropical forests: not seeing the soil for the trees?

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Abstract

Differing perceptions of the impacts on hydrological functions of tropical forest clearance and conversion to other land uses have given rise to growing and often heated debate about directions of public environmental policy in southeast Asia. In order to help bring more balance and clarity to such debate, this paper reviews a wide range of available scientific evidence with respect to the influence exerted by the presence or absence of a good forest cover on regional climate (rainfall), total and seasonal water yield (floods, low flows), as well as on different forms of erosion and catchment sediment yield under humid tropical conditions in general and in southeast Asia in particular. It is concluded that effects of forest disturbance and conversion on rainfall will be smaller than the average decrease of 8% predicted for a complete conversion to grassland in southeast Asia because the radiative properties of secondary regrowth quickly resemble those of the original forest again. In addition, under the prevailing 'maritime' climatic conditions, effects of land-cover change on climate can be expected to be less pronounced than those of changes in sea-surface temperatures. Total annual water yield is seen to increase with the percentage of forest biomass removed, with maximum gains in water yield upon total clearing. Actual amounts differ between sites and years due to differences in rainfall and degree of surface disturbance. As long as surface disturbance remains limited, the bulk of the annual increase in water yield occurs as baseflow (low flows), but often rainfall infiltration opportunities are reduced to the extent that groundwater reserves are replenished insufficiently during the rainy season, with strong declines in dry season flows as a result. Although reforestation and soil conservation measures are capable of reducing the enhanced peak flows and stormflows associated with soil degradation, no well-documented case exists where this has also produced a corresponding increase in low flows. To some extent this will reflect the higher water use of the newly planted trees but it cannot be ruled out that soil water storage opportunities may have declined too much as a result of soil erosion during the post-clearing phase for remediation to have a net positive effect. A good plant cover is generally capable of preventing surface erosion and, in the case of a well-developed tree cover, shallow landsliding as well, but more deep-seated (>3 m) slides are determined rather by geological and climatic factors. A survey of over 60 catchment sediment yield studies from southeast Asia demonstrates the very considerable effects of such common forest disturbances as selective logging and clearing for agriculture or plantations, and, above all, urbanisation, mining and road construction. The 'low flow problem' is identified as the single most important 'watershed' issue requiring further research, along with the evaluation of the time lag between upland soil conservation measures and any resulting changes in sediment yield at increasingly large distances downstream. It is recommended to conduct such future work within the context of the traditional paired catchment approach, complemented with process-based measuring and modelling techniques. Finally, more attention should be paid to the underlying geological

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controls of catchment hydrological behaviour when analysing the effect of land use change on (low) flows or sediment production.

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1. Introduction

In their introductory paper to this special issue, Tomich et al. (this volume) described the so-called 'environmental issue cycle' which consists of seven consecutive stages (Winsemius, 1986). During the first three stages there is a growing awareness and public acceptance of the seriousness of a certain environmental problem, with gradually mounting pressure for action by the responsible authorities. Since this may challenge the effectiveness of existing government policies on the issue, this is usually followed by a debate on the validity of the available evidence on causes and effects. Once a cause and effect chain has been established equivocally in stage 4, options for mitigation of the problem can be considered, negotiated and implemented during the remaining three stages. Naturally, the latter half of the environmental issue cycle hinges on a decisive outcome of such a debate. It is quite possible, however, that the process comes to a halt because of perceived gaps in the understanding of the problem or the quantification of its impacts. Different interest groups may then apply evidence selectively and advocate a position servicing their own interests (Tomich et al., this volume). The environmental impacts of tropical forest clearance and conversion to other land uses, particularly the effect on low flows, represent a case in point, with seemingly mutually excluding viewpoints being expressed not only by different interest groups but also by different representatives of the scientific community. The 'traditional' stance is aptly summarised by the following excerpt from Valdiya and Bartarya (1989) with respect to environmental conditions prevailing in the Kumaun Himalaya in northern India:

> 'The increasingly greater use of land resources has a telling effect on the quality of life of the people of the region. Springs are drying up or becoming seasonal, and the difference in the volume of water flowing down the rivers during the dry and rainy seasons is commonly more than 1000 times, resulting in the too-little-and-too-much-water syndrome, a common feature of desert country. The evaporation of moisture in the soil on the tree-less slopes is very high, and xerophytes (cacti) are beginning to take foothold on naked slopes. These features could be described as harbingers of the onset of desertification'.
>
> The authors go on to say that:
>
> 'It is reasonable to attribute the change in the conditions of weather as reflected in the decline of rainfall and moisture deficiency in the soil to the deterioration—and decimation in some parts—of the forests. For the forests are known to have great influence on rainfall'.

Similarly, there is a widespread belief that logging of upland forested catchments is the root cause of floods occurring in the lowlands and that flood damage can be eliminated by large-scale reforestation. For example, in the words of Sharp and Sharp (1982): 'Overlogging is now officially recognised as the cause of the July 1981 severe flooding of the Yangtze' in China. Similar statements were issued after the devastating floods of 1998 in the same area. A central concept in the 'traditional' view of the role of forests is the 'sponge' effect of the tree roots, forest litter and soil. It has been claimed even that the roots (sic!) soak up water during wet periods and release it slowly and evenly during the dry season to maintain water supplies (Spears, 1982; Myers, 1983). With this in mind, planting trees in degraded areas is often expected to restore the reliability of streams (Eckholm, 1976; Sharp and Sharp, 1982; Nooteboom, 1987; Bartarya, 1989).

The traditional line of thinking on the hydrological role of forest came under scrutiny in the early 1980s when L.S. Hamilton and others began to question the validity of some of the underlying assumptions.

A heated battle on the hydrological role of forest was fought on the pages of the forestry journal of the former Dutch East Indies (*Tectona*) some 60–70 years ago. Protagonists of the ‘sponge’ theory (Steup, 1927; Oosterling, 1927) vigorously opposed the ‘infiltration theory’ (which stated that baseflow is governed predominantly by geological substrate rather than by the presence or absence of a forest cover; Roessel, 1927, 1928, 1939a,b,c; Zwart, 1927). Others (De Haan, 1933; Coster, 1938; Heringa, 1939) took an intermediate position, emphasising the positive influence of forests with respect to the prevention of soil erosion and floods rather than on dry season flows (see Chin A Tam, 1993 for the English abstracts of these papers which were originally written in Dutch). A paired catchment experiment was set up in West Java in 1931 to study the long-term effects of forest clearing for rainfed agriculture on the flows of water and sediment and so settle the debate (De Haan, 1933) but most of the experimental results were lost during World War II, thus illustrating how the ‘environmental issue cycle’ may remain stuck in the debating phase for many years.

In a classic contribution which may be seen as the beginning of a new and more ‘scientific’ view of tropical forest functioning, Hamilton and King (1983) considered that ‘roots may be more appropriately labelled a pump rather than a sponge’ and that ‘roots certainly do not release water in the dry season but rather remove it from the soil in order that the trees may transpire and grow’. Similarly, they considered that ‘major floods occur because too much rain falls in too short a time, or over too long a time. In either case, the rainfall exceeds the capacity of the soil mantle to store it and the stream channel to convey it’. Because of the paucity of hard quantitative evidence from the tropics proper at the time, many of Hamilton’s statements had to be based on research results from the temperate zone (notably the US, New Zealand, Australia, and South Africa) and professional judgement. Nevertheless, his contentions were confirmed later by a series of in-depth reviews of various aspects of the tropical literature by the present author (Bruijnzeel, 1986, 1989, 1990, 1992, 1997, 1998, 2002a; Bruijnzeel and Proctor, 1995; Bruijnzeel and Veneklaas, 1998). The early reviews by Hamilton and King (1983) and Bruijnzeel (1986) were criticised, in turn, by some who were afraid that these views would lead to a ‘selling out to the enemy’ (Smiet, 1987; Nooteboom, 1987). However, as pointed out by Hamilton (1987b), the ‘attackers’ of the traditional line of thought merely aimed at greater accuracy and realism. Ironically, and perhaps partly as a result of the sometimes rather provocative style used by the early messengers of the new line of thought, there seems to be a growing tendency nowadays to emphasise the more ‘negative’ aspects of forests, such as their higher water use and their inability to prevent extreme floods rather than their protective values (enhanced water quality, moderation of most peak flows, carbon sequestration) (Forsyth, 1996; Calder, 1999, 2002; cf. Van Noordwijk et al., this volume). As will be discussed in more detail in the section on stormflows and floods it is important to distinguish between the effects of land cover (‘vegetation’) per se and those of soil water storage capacity.

The aim of the present paper is to review the available evidence with respect to the influence of the presence or absence of a good forest cover on rainfall, streamflow totals and seasonal distribution (peak flows, low flows), as well as on erosion and catchment sediment yields in the humid tropics. Although what follows below draws to a fair extent on two earlier literature reviews by the author (Bruijnzeel, 1993, 1996), an effort has been made to update these publications and to highlight soil and geological aspects. Furthermore, particular attention is paid to southeast Asia, in keeping with the regional focus of the Chiangmai workshop and to answer some (though not all) questions raised in the introductory paper by Tomich et al. (this volume). The paper concludes with various suggestions as to what the author perceives as the most pressing research needs with respect to the hydrological role of forests in southeast Asia and elsewhere in the humid tropics.

2. Tropical forest and precipitation

Although the higher evapotranspiration and greater aerodynamic roughness of forests compared to pasture and agricultural crops will lead to increased atmospheric humidity and moisture convergence, and thus to higher probabilities of cloud formation and rainfall generation (André et al., 1989; Pielke et al., 1998), early reviewers of the subject of forests and rainfall concluded that there was no significant effect. Any observations of enhanced rainfall in forested areas were attributed either to orographic effects (forests being found in uplands where chances of cloud formation were simply greater because of atmospheric cooling of rising air) or to differences in rain gauge exposure to wind and rain (sheltered in forest clearings versus exposed in cleared terrain) (e.g. Penman, 1963; Pereira, 1989).

The discussion is not made any easier by the fact that rainfall in the tropics is notoriously variable, both in space and time (Nieuwolt, 1977; Manton and Bonell, 1993). In addition, irregular cyclic patterns occur, such as the quasi-biennial oscillation (QBO) and the El Niño southern oscillation (ENSO), which exhibit cycles of 2–2.5 years (Parthasarathy and Dhar, 1976; Lhomme, 1981) and ca. 3–8 years (World Meteorological Organization, 1988), respectively. Moreover, at a somewhat longer time scale, it has been suggested that cyclic patterns in rainfall of about 10, 21 and 32 years are related to variations in single, double and triple sunspot activity, respectively (Vines, 1986). More recently, variability in cross-equatorial heat transport by ocean surface currents was identified as a major factor contributing to temporal instabilities in monsoon systems (Zahn, 1994). Clearly, as long as our understanding of the complex interactions between factors influencing natural climatic variability remains limited, it will be very difficult, if not impossible, to separate man-made impacts on climate (e.g. through 'deforestation') from natural variability (Mahé and Citeau, 1993; Street-Perrott, 1994). Fortunately, however, considerable progress has been made in this respect in recent years using atmospheric circulation models.

Shukla (1998) demonstrated that tropical atmospheric flow paths and rainfall, especially over the open ocean and in the more 'maritime' tropics, are so strongly determined by the underlying sea-surface temperature (SST) that they show little sensitivity to changes in initial conditions of the atmosphere (humid or dry). Shukla further hypothesised that the latitudinal dependence of the rotational force of the earth and solar heating together produced the unique structure of the large-scale tropical motion field, such that, for a given boundary condition of SST, the atmosphere is stable with respect to internal changes. As a result, it is now quite possible, once an ENSO event has begun, to predict its growth and maturation for the following 6–9 months (Shukla, 1998). Furthermore, global-scale simulations by Koster et al. (2000) showed that land and ocean processes have rather different domains of influence in the world. The amplification of precipitation variance by land–atmosphere feedbacks appears to be most important outside of the (tropical) regions that are affected most by SST. In other words, the impact of land cover on the precipitation signal is expected to be muted in regions with a large oceanic contribution, such as southeast Asia and the Pacific, West Africa, the Caribbean side of Central America and northwestern South America (Koster et al., 2000).

Two basic approaches are usually followed to study the effects of land-cover change on rainfall: (i) trend analysis of long-term rainfall records in combination with concurrent information on changes in land use; and (ii) computer simulation of regional (or global) climates. Not surprisingly, model simulations have been on the increase in recent years and, as hinted at already in the previous paragraphs, they have contributed to an improved understanding of the respective factors and mechanisms. In the following, results obtained with each of the two approaches are summarised.

2.1. Time series analysis

Circumstantial evidence for (at least temporarily) decreased rainfall at individual or groups of rainfall stations in the tropics abounds in the literature (see review by Meher-Homji, 1989). Although such reports often blame 'deforestation' (e.g. Valdiya and Bartarya, 1989), they have rarely taken into account large-scale weather patterns (including ENSO events) or the above-mentioned cyclical fluctuations, nor have they applied rigorous statistical techniques. Those investigators who did, usually found trends to be either non-existent or non-significant, or at best only weakly significant (e.g. Mooley and Parthasarathy, 1983 for 306 station records between 1871 and 1980 in India; Fleming, 1986 for 10 station records of up to 95 years duration in Costa Rica; Oyo, 1987 for 60 station records between 1910 and 1985 in West Africa). Tangtham and Sutthipibul (1989) reported a significant negative correlation between 10-year moving averages of annual rainfall and remaining forest area in northern Thailand for the period 1951–1984 whilst a positive correlation was found between forest area and the number of rainy days. However, the authors pointed out that the effect of deforestation, if any, was still within one standard error of the means for the respective time series. Indeed, Wilk et al. (2001) were unable to detect any widespread changes in rainfall totals or patterns in the 12,100 km^2 Nam Pong basin in northeast Thailand between 1957 and 1995, despite a reduction in the area classified as forest from 80 to 27% during the last three decades. Nevertheless, rainfall in the whole of Thailand shows a remarkable decreasing trend since the 1950s during the month of September, i.e. when the southwest monsoon current is weakening. In July and August, when the monsoon is still strong, no such decrease is noted (Yasunari, 2002). A related atmospheric modelling study by Kanae et al. (2001) suggested that this decrease in September rainfall may at least partly be related to changes in surface albedo and roughness due to deforestation. Like Wilk et al. in Thailand, and working on a still larger scale, Costa et al. (2003) did not find any effect on rainfall totals or distribution following the conversion of cerrado vegetation (shrubs and scattered trees) to pasture over 19% (ca. 33,000 km^2) of the Tocantins River basin in sub-humid (1600 mm per year) east-central Brazil.

Several authors noted that drought conditions in West Africa were particularly persistent between the mid 1960s and early 1990s, after which rainfall increased again (Oyo, 1987; Mann, 1989; Adejuwon et al., 1990; Olivry et al., 1993; Zeng et al., 1999). Although some have blamed 'deforestation' in this respect (e.g. Mann, 1989), work by Mahé and Citeau (1993) has shown that this persistent dry period corresponded with the occurrence of SST anomalies over the Atlantic ocean. A strong oceanic influence on Sahelian rainfall was also found during global (Koster et al., 2000) and regional (Dolman et al., 2004) atmospheric modelling exercises. However, although incorporation of land-surface characteristics (albedo, soil moisture status) in an atmosphere–ocean circulation model did not improve the correlation between observed and predicted year-to-year rainfall variability, substantial improvement was obtained for inter-decadal (>10 years) rainfall variability (Zeng et al., 1999; Fig. 1). Zeng et al. (1999) ascribed the lack of influence exerted by land-surface feedbacks at the short term to the existence of a phase lag between the occurrence of rainfall and the adjustment (recovery) of the vegetation.

Despite the problem of interannual variability, evidence of persistent trends in either the total amount of rainfall or the intensity of the dry season, or both, is accumulating for various parts of Asia. For example, Wasser and Harger (1992) reported that the average length of the dry season at five (urban) stations in Java had increased from 4.4 months at the beginning of the 20th century to 5.4 months in 1991. The slope of this trend differed from zero at the 1% significance level. Wisely, the authors did not touch upon the question as to whether the inferred tendency towards increased aridity was related to changes in land use (such as the gradual urbanisation of the island; Whitten et al., 1996). However, they did note that drought years prior to 1970 were not always related to ENSO events whereas the seven droughts occurring between 1970 and 1991 were. Although this seems to suggest an external rather than a 'local' cause (cf. Shukla, 1998; Koster et al., 2000), the wealth of climatic information for the Indonesian archipelago (more than 4450 rainfall stations in 1941 versus about 2900 in 1988, with many records spanning more than 120 years, especially in Java; Berlage, 1949; Van der Weert, 1994) invites further in-depth analyses of records of rainfall and land-cover change. However, bearing in mind the conclusions of Shukla (1998) and Koster et al. (2000) that under the 'maritime' tropical conditions prevailing in Indonesia oceanic influences are likely to dominate rainfall variability, a combination of land-cover time series analysis and meso-scale atmospheric modelling would seem to be the most promising way forward here. Another long-term negative trend in rainfall (of about 5.5 mm per year on average since the late 19th century) has been described for upland southern Sri Lanka by Madduma Bandara and Kuruppuarachchi (1988). Although the area in question experienced substantial conversion of (montane) forest to tea plantations, it remains to be seen whether the associated drop in evapotranspiration (which is likely to be modest; see section on total water yield) over such a limited area (<500 km^2) would be sufficient to affect regional rainfall. Alternatively, the change in rainfall may be related to a shift in the location of the intertropical convergence zone (equatorial trough) over Sri Lanka, reflecting larger-scale changes in atmospheric circulation (Arulanantham, 1982). Further work is needed which

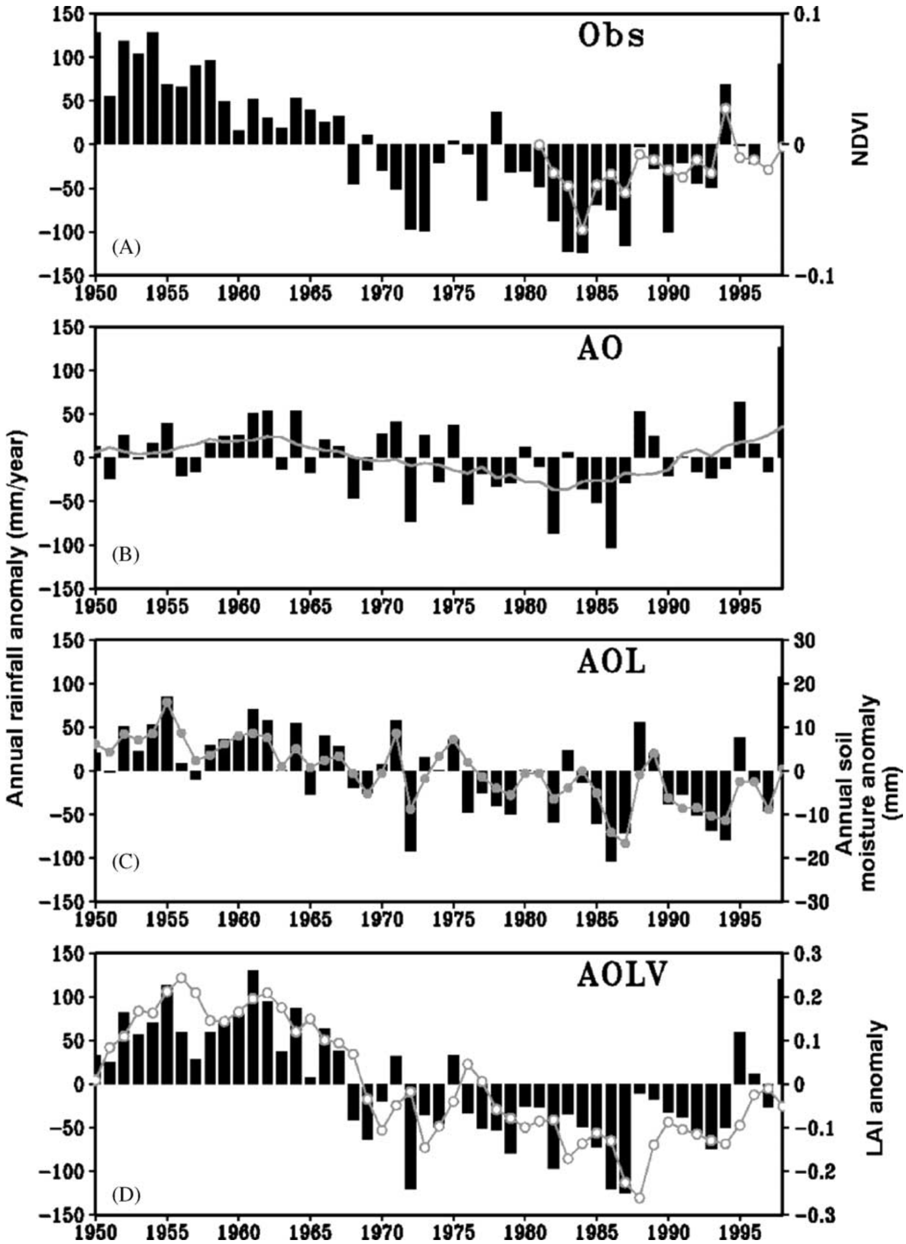

Fig. 1. Annual rainfall anomaly (vertical bars) over the West African Sahel (13–20°N, 15°W–20°E) from 1950 to 1998: (A) observations; (B) model with non-interactive land-surface hydrology (fixed soil moisture) and non-interactive vegetation (SST influence only; AO). Smoothed line is a 9-year running mean showing the low-frequency variation. (C) Model with interactive soil moisture but non-interactive vegetation (AOL); (D) model with interactive soil moisture and vegetation (AOLV) (after Zeng et al., 1999).

could usefully employ an atmospheric circulation model. Fu (2002) did just that to unravel the causes of the significant increase in aridity index over East China since the 1880s. The atmospheric moisture situation over East China is mainly related to the intensity of the summer monsoon and a gradual drying would imply a corresponding weakening of the summer monsoon. Since large parts of the region were converted to rainfed agriculture a meso-scale atmospheric model was combined with a land-surface parameterisation scheme to simulate the potential impacts of land-cover change. Indeed, the model predicted a weakening of the summer monsoon as a result of changes in surface roughness, leaf area index and reflection coefficient. Thus, the observational evidence concurs with model predictions in suggesting that large-scale land-cover change in East Asia is indeed capable of producing changes in the regional surface climate (Fu, 2002).

2.2. Simulation studies

In view of the perceived role of, especially, the Amazon forest in the regulation of the regional climate (Salati and Vose, 1984) a number of increasingly sophisticated computer simulations have been carried out since the mid-1970s to assess the climatic consequences of a large-scale forest conversion to pasture. The results of the respective modelling efforts (reviewed by Henderson-Sellers et al., 1993; McGuffie et al., 1995; Lean et al., 1996; Costa, 2004) vary considerably, depending, inter alia, on the land-surface parameterisation scheme used. However, whilst the magnitude of the predicted changes in surface temperature, evaporation and precipitation following forest conversion may differ between simulations, there is a growing consensus that temperatures will increase whereas both evaporation and rainfall will be reduced (Henderson-Sellers et al., 1993; McGuffie et al., 1995). Such findings reflect the lower water use and aerodynamic roughness of pasture compared with forest, and thus the degree of atmospheric moisture convergence and turbulence which, ultimately, affect cloud formation and rainfall generation (Pielke et al., 1998). On the other hand, as pointed out by Eltahir and Bras (1993) and Costa (2004), there is less agreement for the predicted change in runoff (derived as the difference between the change in rainfall minus that in evaporation). Some models have predicted an increase in runoff and others a decrease. Interestingly, as noted by Bruijnzeel (1996), the magnitude of the predicted changes in rainfall, etc. seems to become smaller as the models become more refined and the parameterisation of the land surface improved. One of the more sophisticated of these simulations (Lean et al., 1996) derived an average increase in temperature of 2.3 °C and a reduction in annual rainfall of 7% (0.43 mm per day or ca. 150 mm per year). Conversely, increases in rainfall were predicted for the outer parts of the basin (Colombia, Ecuador and Perú), and to a lesser extent for the southern rim (cf. Chu et al., 1994). It should be noted, however, that this particular modelling exercise involved a five-fold reduction in soil infiltration capacity after conversion to pasture. This led to a much increased production of surface runoff and thus to diminished soil water reserves which, in turn, limited water uptake by the grassland. When the surface intake capacity of the soil was maintained at the former level the model produced a somewhat smaller reduction in rainfall (0.3 mm per day or ca. 110 mm per year; Lean et al., 1996). Such comparatively small reductions in rainfall after conversion appear to challenge the widely accepted notion that as much as 50% of the rainfall over Amazonia is generated by the forest itself (Salati and Vose, 1984), perhaps because the degree to which the Amazon Basin represents a closed system has been overestimated in the past (Eltahir and Bras, 1994).

Dirmeyer and Shukla (1994) demonstrated that the significant decreases in precipitation predicted in most Amazonian deforestation simulations depended most strongly on the prescribed shift in the reflection of short-wave radiation (albedo) when going from forest to pasture. Albedo changes of +0.08 have typically been used in these experiments but the albedo of existing deforested areas in Amazonia has been shown to be only 0.03–0.04 higher than that of undisturbed forest, mainly because secondary successional vegetation rather than pasture tends to become the dominant land cover. In reality, therefore, changes in temperature, evaporation and precipitation will be smaller than predicted for the extreme case represented by the simulations (Giambelluca, 1996; Sommer et al., 2002; cf. Costa et al., 2003). Also, Bonell and Balek (1993) made the pertinent point that insufficient attention has been given to the adequate parameterisation of lateral surface flows and soil hydrological aspects (including catchment water yield; cf. Van Noordwijk et al., this volume). Indeed, the fact that Lean et al. (1996) expressed surprise upon finding that changes in their simulated evaporation and rainfall figures were strongly influenced by the value used for the infiltration capacity of the pasture soil already indicates the need for catchment process hydrologists and climate modellers to work more closely together (cf. Nemec, 1994; Bonell, 1998). It also suggests that future model results may differ from those obtained with the present generation of models.

Despite these caveats there is reason for concern. Of late there is increasing observational (as opposed to purely model-based) evidence that forest conversion over areas between 1000 and 10,000 km^2 causes feedbacks in the timing and spatial distribution of clouds. For example, Cutrim et al. (1995) documented how development of clouds occurred later during the day over deforested parts of southwest Amazonia. Similarly, using satellite imagery, Lawton et al. (2001) demonstrated substantially reduced cloud formation over the deforested parts of the Atlantic coastal plain in northern Costa Rica during the dry season. Further north, in Nicaragua where a good forest cover is still maintained, there was no such reduction. Lawton et al. ascribed this contrast to differences in energy partition between forest and pasture and they were able to reproduce their observations using (a more or less inverse application of) the RAMS meso-scale atmospheric circulation model (Pielke et al., 1992). However, not only did Lawton et al. have to use parameter values derived for central Amazonian forest and pasture, they also applied a rather arbitrary contrast in soil water contents below forest and pasture (much higher under forest). Generally speaking, caution is needed when interpreting model results obtained with uncalibrated land-surface parameterisation schemes or parameter values derived at locations with rather contrasting climatic and soil conditions (Dolman et al., 2004). An interesting recent finding which may offer a potential alternative explanation for reduced cloud formation above deforested areas is that biogenic aerosols produced by large areas of forest appear to play an important role as cloud condensation nuclei during convection (Roberts et al., 2001; Silva Dias et al., 2002). Interestingly, not all types of large-scale rain forest conversion to agricultural cropping would seem to have such a negative climatic impact. Yasunari (2002) relates how on the vast central and southern China plain more than 80% of the area is occupied by irrigated rice fields during the rainy season. A meso-model study using measured surface energy and water fluxes was conducted for two contrasting situations: one with irrigated rice fields (where evaporation exceeds the sensible heat flux), and another with rainfed farmland conditions (where sensible heat flux exceeds evaporation). The experiment showed a dramatic contrast in moisture content of the atmospheric boundary layer (ABL) associated with the two land uses. Above irrigated rice the ABL was much wetter and deep convection (cloud formation) and thus strong rainfall, developed much more easily. Such results suggest that the rice paddies of east and southeast Asia may well represent a form of land use that is in harmony with the prevailing monsoonal climate through its positive atmospheric feedback (Yasunari, 2002). Other important hydrological benefits of irrigated rice cultivation include delaying the arrival of surface runoff peaks and trapping sediment eroded further upslope (Purwanto, 1999).

In contrast to using parameter values derived for forest and pasture in Amazonia, Van der Molen (2002) made detailed micro-meteorological measurements above a coastal wetland forest and well-watered pasture in the northern coastal plain of Puerto Rico to study the climatic effects of forest conversion. Evaporation from the forest was about 14% lower than that from the pasture while the sensible heat flux (warming up of the overlying air) was about twice as high. These findings differ markedly from the results obtained in Amazonia cited earlier in this section (Lean et al., 1996), possibly because of the presence of brackish groundwater in the Puerto Rican forest. Next, Van der Molen used his measurements to calibrate the land-surface model within RAMS and using the full three-dimensional set-up of the model he investigated the effect of a complete conversion of coastal plain wetland forest to pasture on cloud formation in the adjacent Luquillo Mountains. The partitioning of energy above the original forest produced a warmer and slightly less moist boundary layer and because of this greater thermal contrast between ocean and land, the sea breeze was stronger than in the grassland case. The associated stronger updrafts tended to bring the moisture to higher elevations, thereby increasing the possibility of generating clouds. After conversion to pasture the sea breeze effect was decreased. Interestingly, six out of eight rainfall stations in the lowlands of Puerto Rico exhibit a statistically significant negative long-term trend in rainfall (Van der Molen, 2002). Dolman et al. (2004) suggested that a similar scenario might also apply to the Costa Rican case described above and so offer an alternative explanation for the observed increase in cloud cover above lowland forest there. This is less likely, however, because of the absence of brackish groundwater in all but a narrow coastal strip and the relatively low relief in the area studied by Lawton et al. (2001). Indeed, Nooteboom (1987) related how the diurnal cycle of land-sea breezes appeared to become more intense after the widespread replacement of lowland forest by *Imperata* grasslands in southeastern Kalimantan. The above examples serve to illustrate the influence of land cover on atmospheric processes such as cloud formation and possibly rainfall at the meso-scale, also under more 'maritime' conditions (Puerto Rico, eastern Costa Rica). This arguably defies the contention of Shukla (1998) and Koster et al. (2000) that SST is the dominant causative factor under such conditions. Larger-scale conversions (>100,000, >1,000,000 km^2) may cause even more pronounced changes in atmospheric circulation, to the extent of actually affecting precipitation patterns, even under more continental climatic conditions, such as in Amazonia. In a recent 'interactive' (i.e. allowing atmospheric circulation feedbacks) simulation of the hydrological impact of large-scale Amazonian forest conversion to pasture, Costa and Foley (2000) demonstrated how inclusion of such feedbacks resulted in the prediction of larger declines in evapotranspiration and, especially, rainfall than in the case without feedbacks. As a result, runoff changed from a predicted increase of 0.5 mm per day (no feedbacks) to a decrease of 0.1 mm per day. It would seem, therefore, that the normally observed increase in streamflow totals after forest clearing at the local scale (see section on water yield below) may be moderated or even reversed at the largest scale because of the concomitant reduction in rainfall induced by atmospheric circulation feedbacks (Costa, 2004).

Although some of the simulations of the effects of large-scale deforestation (conversion to grassland) on climate have included Africa or southeast Asia as well (e.g. Polcher and Laval, 1994; Henderson-Sellers et al., 1996), the parameterisations used in these simulations also drew heavily on data collected in central Amazonia. Once again, the predicted effects are likely to be (much?) less severe than the 8% reduction in rainfall derived by Henderson-Sellers et al. (1996) for southeast Asia because the actual parameter settings representative of the secondary vegetation types replacing the forest resemble those of the forest much more than the more extreme grassland scenario used in the simulation (Giambelluca et al., 1996, 1999). There are indications, however, that both total forest evapotranspiration and rainfall interception under the more 'maritime' tropical conditions prevailing in the region may be higher than those determined in central Amazonia (Schellekens et al., 2000). Needless to say, this may have important ramifications for the outcome of simulation studies dealing with the climatic effects of land-cover change. Further work is needed to elucidate such effects.

2.3. Tropical montane 'cloud forests'

Although forests may not directly determine the amounts of precipitation they receive when occurring as scattered stands of limited areal extent, there are specific locations, such as coastal and montane fog or cloud belts, where the presence of tall vegetation may increase the amount of water reaching the forest floor as canopy drip. This is effected via the process of fog or cloud interception, i.e. the capturing of atmospheric moisture by the canopy of these 'cloud forests' where subjected to more or less persistent wind-driven fog or clouds (Zadroga, 1981). Contributions by cloud water interception generally lie within the range of 5–20% of ordinary rainfall at wet tropical locations (Bruijnzeel and Proctor, 1995; Bruijnzeel, 2002a) but can be much higher (>1000 mm per year) at certain particularly exposed locations (Stadtmüller and Agudelo, 1990) although it is not always certain to what extent such high values include wind-driven rain (Cavelier et al., 1996; Clark et al., 1998). Relative values of cloud water interception may also exceed rainfall during the 'dry' season in more seasonal climates (Vogelmann, 1973; Cavelier and Goldstein, 1989; Brown et al., 1996).

Cloud forests seem particularly vulnerable to global warming as they often occur on exposed ridges and mountain tops with shallow soils of limited storage capacity (Stadtmüller, 1987; Werner, 1988). Pounds et al. (1999) recently reported how only slight decreases in the number of days without measurable precipitation (taken as an index of fog frequency) had a profound effect on the composition and size of frog and lizard populations in a (leeward) Costa Rican cloud forest. The most dramatic decreases in population numbers within an overall downward trend occurred during years which had demonstrably higher SST in the Pacific Ocean and presumably reduced fog incidence (Pounds et al., 1999). Such findings illustrate the close link between forest hydrometeorology and biodiversity under the extreme climatic conditions prevailing in the cloud belts of wet tropical mountains (cf. Bruijnzeel and Veneklaas, 1998). A further sign on the wall in this respect relates to the decline in cumulus cloud cover during the dry season over deforested areas in northeastern Costa Rica referred to earlier (Lawton et al., 2001). A regional climatic modelling exercise by the same authors predicted a significant warming of the air above such deforested (pasture) areas to the extent that the average cloud base in the adjacent mountains was lifted significantly (cf. Still et al., 1999; see also the caveats expressed in the previous section in relation to this study). Similarly, the average cloud base in the Luquillo Mountains of eastern Puerto Rico was seen to rise for several months after hurricane ‘Hugo’ had effectively defoliated large tracts of rain forest lower down on the mountain (as opposed to the coastal wetland forest studied by Van der Molen, 2002) in September 1989. The increase in air temperatures associated with the temporarily reduced evaporating capacity of the forest caused the air to condensate at a higher elevation, thereby exposing summit cloud forests that are normally enshrouded in clouds (Scatena and Larsen, 1991; see also photographs in Bruijnzeel and Hamilton, 2000). The effect gradually disappeared as the leaves grew back again after several months (F.N. Scatena, personal communication). In the same area, Scatena (1998) interpreted the presence of isolated stands of large and very old (>600 years) Colorado trees (*Cyrilla racemiflora*) at elevations well below the current cloud base that experience relatively low rainfall (<3000 mm per year) as evidence of a gradual upward shift in vegetation zonation over the past several centuries. *Cyrilla* is currently a dominant tree in areas above the local cloud base (>600 m) and is most common where mean annual rainfall exceeds 4000 mm. Similarly, Brown et al. (1996) reported the occurrence of pockets of mossy cloud forest below the current average cloud base in Honduras. There is a need for more systematic research linking such empirical evidence to records of current and sub-recent climatic change (cf. Scatena, 1998). On single mountains, a lifting of the average cloud condensation level will result in the gradual shrinking of the cloud-affected zone. On multiple-peaked mountains, however, the effect may be not only that, but one of increased habitat fragmentation as well, adding a further difficulty to the chances of survival of the remaining species (Sperling, 2000). Apart from amphibians (Pounds et al., 1999), the epiphyte communities living in the more exposed parts of cloud forest canopies might prove to be equally suited to detecting changes in climatic conditions, and possibly enhanced ozone and UV-B levels as well (Lugo and Scatena, 1992; Benzing, 1998).

Although particularly common in Central and South America, montane cloud forests are also widespread in southeast Asia and the Pacific, occurring at elevations as low as 500–700 m on small oceanic islands to >2000 m on larger mountains (Hamilton et al., 1995; Bruijnzeel, 2002a). As will be shown in the next section, cloud forests are of great hydrological significance in Central America. The few observations available for southeast Asian cloud forests (Bruijnzeel et al., 1993; Kitayama, 1995) suggest modest rates of cloud water interception and low to very low water use. More work is needed to assess the hydrological significance of cloud forests in southeast Asia.

3. Tropical forest and water yield

3.1. ‘Contradictory’ results

As indicated in Section 1, a common notion about the hydrological role of forests is that the complex of forest soil, roots and litter acts as a ‘sponge’ soaking up water during rainy spells and releasing it evenly during dry periods. Upon clearing, the ‘sponge effect’ is lost through the rapid oxidation of soil organic matter, compaction by machinery or grazing, etc. (Lal, 1987), with diminished water yield as a result. Indeed, accounts of springs and streams drying up during the dry season after tropical forest removal are numerous enough (Hamilton and King, 1983; Valdiya and Bartarya, 1989; Pereira, 1989; cf. Pattanayak, this volume). On the other hand, the number of reports of the reverse (i.e. streams drying up in the dry season after reforestation of degraded land) is increasing as well (see below). When trying to reconcile these apparent contradictions, it is helpful to distinguish between the effect of forest clearing on *total water yield* and on the *seasonal distribution of flows* (Bruijnzeel, 1989).

Before addressing the issue in more detail, a few methodological comments are in order. First of all, simply comparing streamflow totals for catchments with contrasting land use types may produce misleading results because of the possibility of geologically determined differences in catchment groundwater reserves or deep leakage (Roessel, 1927, 1939; Meijerink, 1977; Hardjono, 1980). Catchments underlain by sandstones, limestones, basalts and volcanic tuffs are particularly notorious in this respect (Gonggrijp, 1941b; Davis and De Wiest, 1966). Also, the strong spatial and year-to-year variability of tropical rainfall may compound direct comparisons between catchments or years. For example, despite an almost complete conversion from forest to pasture within a 5-year period (1979–1984), average post-forest annual streamflow (1980–1995) from the 131 km^2 Rio Pejibaye basin in southern Costa Rica was about 320 mm lower than under fully forested conditions (1970–1979), almost certainly due to lower rainfall totals (J. Fallas, personal communication).

Rigorous experimental designs (e.g. the 'paired catchment' technique; Hewlett and Fortson, 1983) or well-calibrated models (e.g. Watson et al., 1999) are needed to overcome such problems. The experimental error of the paired catchment method is such that reductions in forest cover of, say, <20% fail to produce a detectable change in streamflow on small catchments (Bosch and Hewlett, 1982), although this does not necessarily hold for large basins as well (Trimble et al., 1987; Costa et al., 2003). Such controlled experiments are time-consuming and expensive and, consequently, relatively few of them have been conducted in the tropics (Bruijnzeel, 1990; Malmer, 1992; Fritsch, 1993). Fig. 2 summarises the evidence available to date.

In all cases, the removal of more than 33% of forest cover resulted in significant increases in annual streamflow during the first 3 years. Initial gains in water yield after complete forest clearance ranged between 145 and 820 mm per year. In addition, increases in water yield proved to be roughly proportional to the fraction of biomass removed (Fig. 2a). These changes in water yield mainly reflect the different evaporative characteristics of mature tropical forest and (very) young secondary or planted vegetation and to a much lesser extent increases in storm runoff (response to rainfall). Under mature tropical rain forest typically 80–95% of incident rainfall infiltrates into the soil, of which ca. 1000 mm per year is transpired again by the trees when soil moisture is not limiting, whereas the remainder is used to sustain streamflow. As such, the bulk of the increase in flow upon clearing is normally observed in the form of baseflow, as long as the intake capacity of the surface soil is not impaired too much (Bruijnzeel, 1990).

The observed variation in initial response to clearing (Fig. 2a) is considerable and can be explained only partially by differences in rainfall between locations or years (Fig. 2b; see Bosch and Hewlett, 1982; Stednick, 1996 for the results of numerous non-tropical studies). Other factors include: differences in elevation and distance to the coast (affecting evaporation; Bruijnzeel, 1990; Schellekens et al., 2000), catchment steepness and soil depth (both of which govern the residence time of the water and the speed of baseflow recession; Roessel, 1939; Ward and Robinson, 1990), the degree of disturbance of undergrowth and soil by machinery or fire (determining both the water absorption capacity of the soil and the rate of regrowth; Uhl et al., 1988; Kamaruzaman, 1991; Malmer, 1992), and the fertility of the soil (influencing post-clearing plant productivity and water uptake: Brown and Lugo, 1990). Because the relative importance of the respective factors varies between sites, additional process studies are usually required if the results of paired basin experiments (which essentially represent a black box approach) are to be fully understood (Bruijnzeel, 1990, 1996; Malmer, 1992; Bonell and Balek, 1993; Sandström, 1998).

3.2. Changes in water yield during forest regeneration

Generally, the initial increases in total water yield following forest clearing exhibit a more or less irregular decline to pre-clearing levels with time, reflecting the development of the regenerating or newly planted vegetation and year-to-year variability in rainfall.

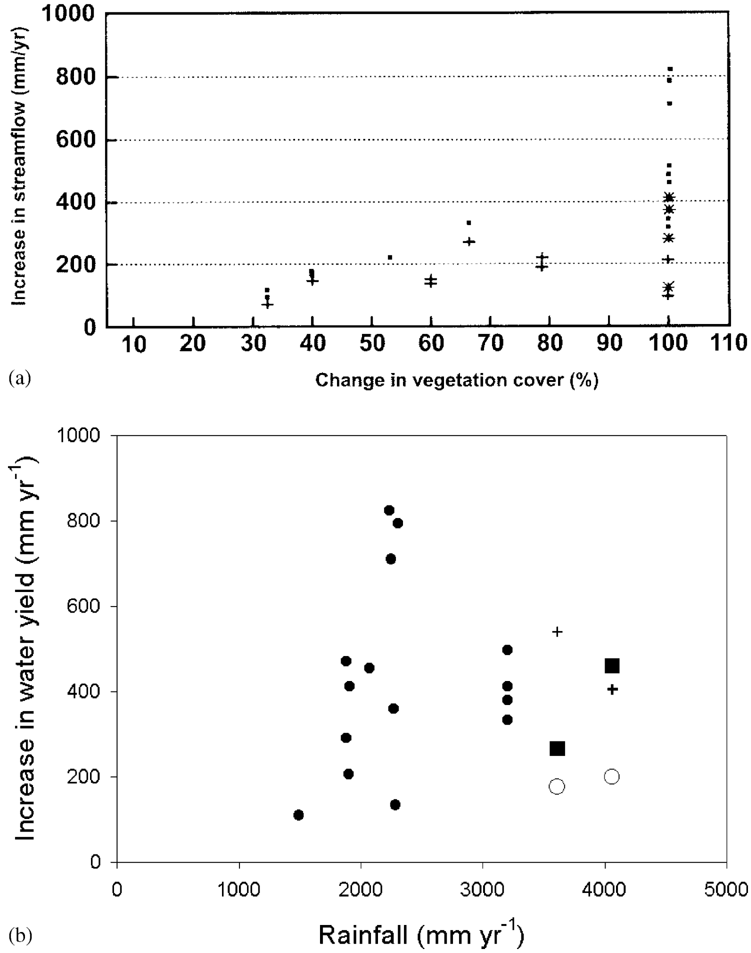

Fig. 2. (a) Increases in annual water yield during the first 3 years after clearing tropical forest vs. percentage of area cleared: (+) dry year, (✱) non-paired catchment study; (b) idem after complete clearing vs. corresponding amounts of annual rainfall ((●) taken from Table 4 in Bruijnzeel (1990), supplemented with data from Malmer (1992) as follows: (+) non-mechanised clearing followed by burning of slash; (○) manual extraction of logs, no burning; (■) mechanised clearing, followed by burning of slash).

Under temperate conditions this may take 3–9 years on shallow soils, depending whether the regeneration is mainly through sprouting or from seeds (Hornbeck et al., 1993) whereas periods of up to 35 years have been reported for regenerating broad-leaved forests on very deep soils (Swank et al., 1988).

A rapid return to pre-disturbance levels of streamflow during forest regeneration after logging or clearing in the humid tropics may be expected in view of the generally vigorous growth of young tropical secondary vegetation (Brown and Lugo, 1990). However, the available information is scarce and to some extent contradictory. Kuraji and Paul (1994) presented results from a water balance study involving two small catchments in Sabah, East Malaysia, which had been subjected to different degrees of forest exploitation two and a half years prior to the observations. One catchment had been selectively logged, the other had been clearfelled and burned. Estimates of forest evapotranspiration (ET) during the third and fourth year after the disturbance were ca. 1450 mm per year for the logged catchment versus ca. 1200 mm per year for the clearfelled basin (where vegetation was dominated by *Macaranga* spp.). Taking these figures at

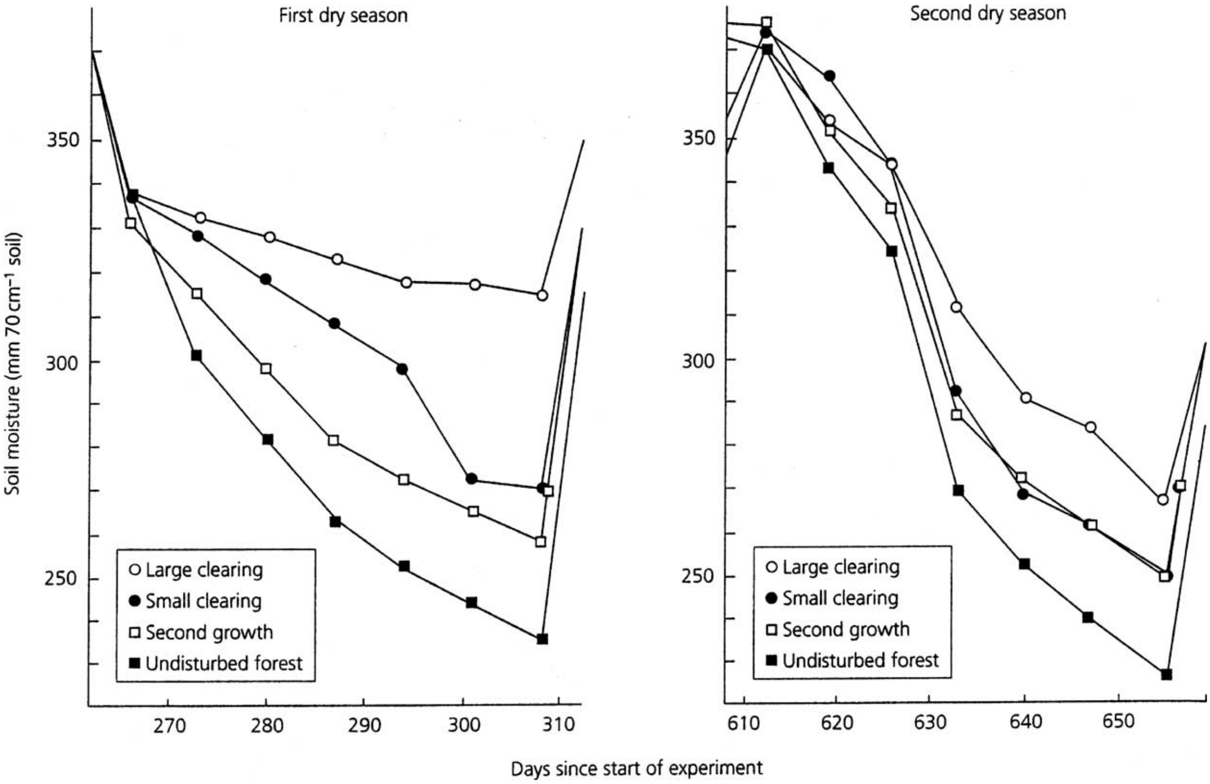

Fig. 3. Soil moisture contents in the top 70 cm of soil below undisturbed forest, 6-year-old secondary growth and in a narrow (10 m × 50 m) and a large (50 m × 50 m) clearing in lowland Costa Rica during two consecutive dry seasons (redrawn after Parker, 1985).

face value, and noting that the estimate for the logged forest is already close to the average value for the ET of undisturbed lowland rain forest in the region (ca. 1465 mm per year; Bruijnzeel, 1990), the water use of young secondary vegetation in southeast Asia may be estimated as being ca. 250 mm per year lower than that for mature forest. Concurrent measurements of above- and below-canopy rainfall in the two forests suggested that this apparent difference in overall ET could be accounted for by the difference in rainfall interception alone (Paul and Kuraji, 1993). Further support for a quick (i.e. within 3–5 years) return of streamflow totals to pre-disturbance levels after logging (but not clearing) may come from the observations of Malmer (1992), also in Sabah. His measurements of streamflow from forested catchments that had lost about one-third of their overstorey biomass 5 years before through logging indicated no trend in annual ET over the next 5 years, except for higher values during wetter years associated with higher rainfall interception losses. Similarly, Parker (1985) found dry season soil water reserves in a small clearing (of similar size as the ones typically created by logging) in Costa Rica to be already indistinguishable from those below 5-year-old regrowth during the second year (Fig. 3). Additional evidence for a rapid recovery of forest water use after clearing and burning comes from eastern Amazonia where the ET of 2-year-old and 3.5-year-old secondary vegetation (determined by micro-meteorological rather than hydrological methods) was estimated at 1365 and 1421 mm per year, respectively (Hölscher et al., 1997; Sommer et al., 2002). The latter values are very similar to that obtained for mature forest in the same area (1350 mm per year; Klinge et al., 2001). Also in eastern Amazonia, Jipp et al. (1998) did not find significant differences in seasonally averaged soil water reserves below mature forest and 15-year-old regrowth at any depth.

Conversely, Abdul Rahim and Zulkifli (1994) did not observe any decline in initial water yield increases (typically 70–100 mm per year) during the first 7 years after harvesting 40% of the commercial stocking in a lowland rain forest in Peninsular Malaysia (the observations were terminated after the seventh year when the area was inundated during the creation of a reservoir). Although this might be taken to suggest that the water use by the regrowth in the gaps created by logging still remained below that of the original stock after seven years, this is less likely in the light of the previous evidence from Amazonia and East Malaysia. Rather, one may think of a more structural cause for the increased flows, such as enhanced runoff contributions by timber extraction roads plus the fact that evaporation from these compacted bare surfaces can be expected to be low. The contradictory result obtained by Abdul Rahim and Zulkifli (1994) once again highlights the need for hydrological process studies supplementing paired catchment experiments (cf. Bruijnzeel, 1996). Hydrological research efforts in secondary vegetation need to be stepped up in general, if only because, with a few exceptions, most tropical countries now have larger areas under secondary vegetation than under primary forest (Brown and Lugo, 1990; Whitmore, 1998; Giambelluca, 2002; see also Hölscher et al., 2004 for a recent summary of the hydrological and soil changes associated with regeneration of tropical forest). Such work should also lead to a better quantitative explanation of the apparent lack of hydrological response to forest alteration in the case of large catchment areas with vegetation in various stages of regeneration (see below).

3.3. Changes in water yield following forest conversion

Whilst streamflow totals are observed to eventually return to pre-clearing levels when regrowth is allowed, the conversion of native tropical forest to other types of land cover may produce permanent changes. For example, permanent increases in annual water yield are usually associated with the conversion of forest to agricultural cropping. Reported increases range from 140 mm per year under the sub-humid conditions prevailing in Nigeria (Lal, 1983) to 410 mm per year in the mountains of Tanzania (Edwards, 1979). The reduced water use of annual crops compared to a full-grown forest reflects not only the diminished capacity of short vegetation to intercept and evaporate rainfall (Van Dijk and Bruijnzeel, 2001b) but also to extract water from deeper soil layers during periods of drought (Eeles, 1979). The former relates primarily to the lesser aerodynamic roughness of short annual crops (and possibly to their smaller leaf area), whereas the reduced water uptake of crops reflects their more limited rooting depth (Calder, 1998; cf. Nepstad et al., 1994). For the same reasons, the conversion of tropical forest to pasture generally produces permanent increases in streamflow as well (150–300 mm per year depending on rainfall; Mumeka, 1986; Fritsch, 1993; Jipp et al., 1998). Similarly sized permanent increases in flow can be expected for forest conversion to plantations of tea (Blackie, 1979a), rubber (Montény et al., 1985) or cocoa (Imbach et al., 1989).

On the other hand, water yields have been reported to return to original levels within 8 years where pine plantations replaced natural forest, such as in upland Kenya (Blackie, 1979b). A similar result may be expected on the basis of the comparable ET values derived for mature oil palm (Foong et al., 1983) and lowland rain forest (Bruijnzeel, 1990), although additional work is needed to test this contention. Bruijnzeel (1997) pointed out how little is known about the hydrological consequences of the planting of such widely used fast-growing tree plantation species as *Acacia mangium*, *Gmelina arborea*, *Paraserianthes falcataria*, and (to a lesser extent) *Eucalyptus* spp. and pines. In view of the very high rainfall interception fractions reported for *A. mangium* in both Peninsular (Lai and Salleh, 1989) and East Malaysia (A. Malmer, personal communication), and the exceptionally rapid growth of this species (Lim, 1988), it is not unthinkable that the total water use of *Acacia* stands may exceed that of the original forest. Recent observations in 10-year-old stands have confirmed the high water use of *A. mangium* in East Malaysia even during a period of drought (Cienciala et al., 2000).

The planting of eucalypts has met particularly vigorous opposition in the popular environmental literature, mainly because they are claimed to be ‘voracious consumers of water’ (e.g. Vandana and Bandyopadhyay, 1983). Plantations of *Eucalyptus camaldulensis* and *E. tereticornis* in southern India indeed exhibited such behaviour with transpiration rates of up to 6 mm per day when unrestricted by soil water deficits at the end of the monsoon, although values fell to 1 mm per day when soil water contents were low during the subsequent long dry season (Roberts and Rosier, 1993). Because of this regulating mechanism the annual water use of the plantations on soils of intermediate depth (ca. 3 m) was not significantly different from that of indigenous dry deciduous forest (Calder et al., 1992). However, on much deeper soils (>8 m) the annual water use of the plantations exceeded annual rainfall considerably, suggesting 'mining' of soil water reserves that had accumulated previously in deeper layers during years of above-average rainfall. Moreover, the rate of root penetration was shown to be at least as rapid as 2.5 m per year and roughly equalled above-ground increases in height (Calder et al., 1997). Similar observations have been made below *E. grandis* in South Africa (Dye, 1996). It is probably pertinent in this respect that Viswanatham et al. (1982) observed strong decreases in streamflow after coppicing of *E. camaldulensis* in northern India. A recent experiment involving *E. globulus* in South India (Sikka et al., 2003) confirmed the enhanced effect of coppicing: the reduction in water yield during the second rotation of 10 years (first generation coppice) was substantially higher (by 156%) compared to that during the first rotation (Samraj et al., 1988). The above findings confirm the original fears of Vandana and Bandyopadhyay (1983). Planting of eucalypts, particularly in sub-humid climates, should therefore be based on judicious planning, i.e. away from water courses and depressions or wherever the roots would have rapid access to groundwater reserves (see also the section on effects of reforestation).

No declines in annual streamflow totals have been reported following lowland tropical forest removal. However, it is possible that the clearing of certain types of tropical montane cloud forests for the cultivation of temperate vegetables or the creation of grazing land presents an exception to the rule. Evapotranspiration in cloud forests is known to be low. This, together with the extra inputs of moisture generated through cloud water interception, causes the associated runoff coefficients to be very high (Bruijnzeel and Proctor, 1995; Zadroga, 1981). Despite the demonstrated importance of these forests for the continued supply of water to adjacent lowlands, they are rapidly converted to agricultural uses in many places, particularly in Latin America (Hamilton et al., 1995; Bruijnzeel and Hamilton, 2000). The associated effect on water yield is as yet unknown, but presumably reflects a trade-off between the loss of the extra water formerly gained via interception of cloud water and wind-driven rain, and the difference in water use between the old and the new vegetation. Amounts of cloud interception may differ strongly between cloud forest type and site exposure, however, and the eventual effect of cloud forest clearing on streamflow will therefore depend on the relative proportions of the catchment occupied by the respective forest types (Bruijnzeel, 2002a). For example, exposed ridge top forests may intercept very large amounts of cloud water and wind-driven rain but their spatial extent is generally too small to have a significant effect at the catchment scale (Weaver, 1972; Brown et al., 1996). Most of the available evidence with respect to the hydrological effect of cloud forest conversion relates to diminished *dry season flows* (see the next section). Ataroff and Rada (2000) recently presented (spot?) measurements of forest and pasture water use in montane Venezuela which, after extrapolation to annual values, suggested that streamflow might well decline following conversion. They further reported consistently lower moisture levels in the top 30 cm of the soil below pasture compared to forest to support this contention. However, the inferred total ET of their lightly grazed pasture (2690 mm per year excluding soil evaporation) must be considered unrealistically high as it almost certainly exceeds amounts of available radiant energy at this elevation (2350 m). Further work is necessary.

3.4. Effects of scale

The results presented thus far pertain mostly to small headwater catchment areas (usually $<1\,km^2$) involving a unilateral change in cover. Although these experiments provide a clear and consistent picture of increased water yield after replacing tall vegetation by a shorter one and vice versa, such effects are often more difficult to discern in larger catchments which usually have a variety of land use types and temporal changes therein. In addition, there are complications wherever rainfall exhibits strong spatial variability and withdrawals of water for municipal, agricultural and industrial purposes are large, such as in many densely populated tropical lowland areas. For example, Qian (1983) was unable to detect any systematic changes in streamflow from catchments ranging in size from 7 to 727 km^2 on the island of Hainan, southern China, despite a 30% reduction in tall forest cover over three decades. Dyhr-Nielsen (1986) and Wilk et al. (2001) arrived at essentially the same conclusion for large (12,100–14,500 km^2) river basins in northern Thailand which had lost at least 50% of their tall forest cover since the 1950s. The prime cause of forest alteration in these examples was shifting cultivation. Therefore, apart from any moderating effects of spatial variability in rainfall, this lack of a clear change in streamflow totals must reflect the rapid return to pre-disturbance values of the evaporative characteristics (notably albedo) of the vegetation, and possibly recovered soil infiltration capacities. Giambelluca et al. (1999) reported that the albedo of 8–25-year-old secondary vegetation in northern Thailand was already similar to or lower than that of mature forest (suggesting similar water use). Also, Fritsch (1993) noted that increases in stormflow (by ca. 30%) during slash and burn cultivation in French Guyana disappeared rapidly once the forest was allowed to regenerate. Lal (1996) demonstrated how five years of fallowing after traditional clearing in Nigeria produced a 10-fold increase in soil infiltration capacity.

On the other hand, Madduma Bandara and Kuruppuarachchi (1988) observed an increase of about 200 mm in averaged annual flow totals for the 1100 km^2 upper Mahaweli catchment in Sri Lanka over the period 1940–1980, despite a weak negative trend in rainfall over the same period. Although both trends were not statistically significant at the 95% significance level, the associated increase in annual runoff ratios was highly significant, whereas dry season flows gradually diminished as well (Fig. 4). The increased hydrological response was ascribed to the widespread conversion of tea plantations (not forest) to annual cropping and home gardens without appropriate soil conservation measures (Madduma Bandara and Kuruppuarachchi, 1988).

However, Elkaduwa and Sakthivadivel (1999) obtained a much less consistent picture when analysing a longer time series (1940–1997) of flows from the nearby 380 km^2 upper Nilwala catchment, which had experienced a 35% reduction in forest cover.

Moving further up in scale, Van der Weert (1994) compared streamflow totals for the 4133 km^2 Citarum river basin in West Java, Indonesia, for the periods 1922–1929 and 1979–1986. Average annual rainfall totals for the two periods were very similar at 2454 and 2470 mm, respectively. The corresponding average streamflow totals were 1137 and 1261 mm, suggesting a decrease in apparent catchment ET of

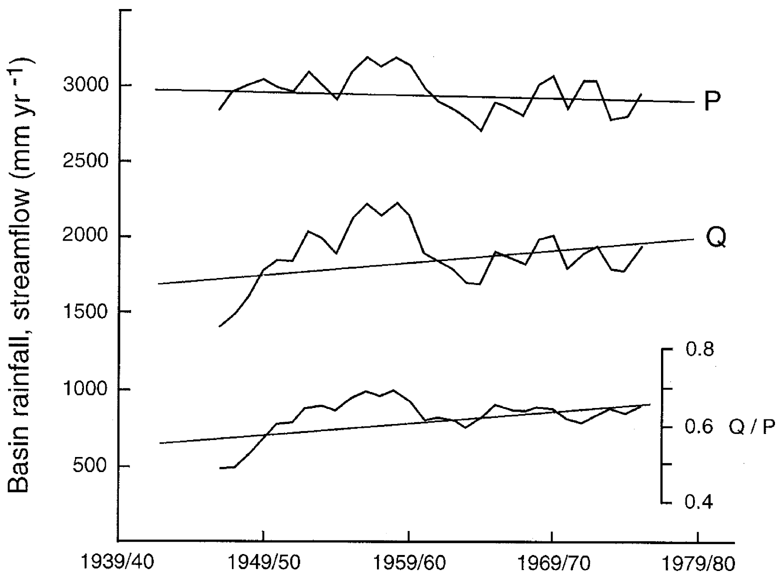

Fig. 4. Five-year moving averages of annual rainfall, streamflow and runoff ratios for the upper Mahaweli basin above Peradeniya, Sri Lanka (after Madduma Bandara and Kuruppuarachchi, 1988).

about 110 mm per year. Although no detailed information is available on pre-war land use in the basin, there was reportedly comparatively little forest clearance. In 1985, almost 50% of the catchment was covered by forest, plantations or mixed gardens, whereas settlements and irrigated rice fields occupied 7 and 34%, respectively, with rainfed fields making up the remaining 9% (Van der Weert, 1994). Because the areas covered by settlements and rice paddies are likely to have increased considerably between the two periods (Whitten et al., 1996), the reduction in overall ET, and therefore the increase in water yield, must be attributed primarily to the increase in areas with compacted surfaces, such as roads and settlements (water consumption by irrigated rice may be as high as that of forest; Wopereis, 1993). Budi Harto and Kondoh (1998) obtained a very similar drop in ET (110 mm) after a 5% increase in settlement area and a 10% increase in rainfed agriculture (both at the expense of irrigated rice fields) elsewhere in West Java. Binn-Ithnin (1988) also reported greatly increased runoff volumes associated with urban growth in the Kuala Lumpur area, Peninsular Malaysia.

A similar cause (i.e. increased urbanisation) may well explain the (rather considerable) increase in annual runoff coefficients derived for several large river basins (25,500–66,625 km^2) in the upper Yangtze valley in southwest China, despite the claim of Cheng (1999) that these increases in flow reflected the 'construction of shelter forests' (which included eucalypts and various phreatophytes) and thus improved infiltration opportunities in only ca. 7% of the basin area. Finally, at the very large scale (175,360 km^2) Costa et al. (2003) relate how an increase of 19% (ca. 33,000 km^2) in the area under pasture at the cost of *cerrado* vegetation in a sub-humid part of Brazil resulted in a significant increase in mean annual discharge (24% or ca. 88 mm per year). Because the change in rainfall was not statistically significant and because the increase in flows was largest during the rainy season, Costa et al. inferred that a reduction in soil infiltration capacity after grazing (cf. Elsenbeer et al., 1999; Godsey and Elsenbeer, 2002) rather than reduced water use by pasture was the chief cause of the observed increase in water yield. Effects of urbanisation and roads must be considered negligible in this particular case. There is a need for increased research efforts to establish the downstream effects of land use change (both positive and negative) at the meso- to macro-catchment scale (see also below).

4. Changes in flow regime following tropical forest conversion

4.1. Dry season flow

In areas with seasonal rainfall, the distribution of streamflow throughout the year is often of greater importance than total annual water yield. As indicated earlier, reports of greatly diminished streamflows during the dry season after tropical forest clearance are numerous. At first sight, this seems to contradict the evidence presented earlier, that forest removal leads to higher overall water yields and wetter soils (Figs. 2–4), even more so because the bulk of the increase in flow after *experimental* clearing is usually observed during conditions of baseflow (Bosch and Hewlett, 1982; Bruijnzeel, 1990). However, the circumstances associated with controlled (short term) catchment experiments may well differ from those of many 'real world' situations *in the longer term*. To begin with, the continued exposure of bare soil after forest clearance to intense rainfall (Lal, 1987, 1996), the compaction of topsoil by machinery (Kamaruzaman, 1991; Malmer and Grip, 1990) or overgrazing (Costales, 1979; Gilmour et al., 1987), the gradual disappearance of soil faunal activity (Aina, 1984; Lal, 1987), and the increases in area occupied by impervious surfaces such as roads and settlements (Rijsdijk and Bruijnzeel, 1990, 1991; Van der Weert, 1994; Ziegler and Giambelluca, 1997), all contribute to gradually reduced rainfall infiltration opportunities in cleared areas. As a result, catchment response to rainfall becomes more pronounced and the increases in storm runoff during the rainy season may become so large as to seriously impair the recharging of the soil and groundwater reserves feedings springs and maintaining baseflow. In other words: the 'sponge effect' is lost. When this critical stage is reached, diminished dry season (or 'minimum') flows inevitably follow (Fig. 5a) despite the fact that the reduced evaporation associated with the removal of forest should have produced higher baseflows. If, on the other hand, soil surface characteristics after clearing are maintained sufficiently to allow

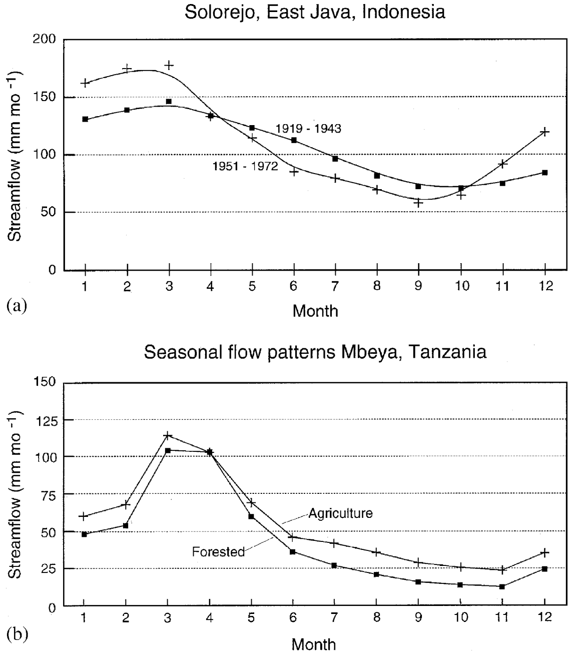

Fig. 5. (a) Change in seasonal distribution of streamflow following changes in land use: replacing 33% of forest by rainfed cropping and settlements in East Java, Indonesia (after Rijsdijk and Bruijnzeel, 1991). (b) Change in seasonal distribution of streamflow following changes in land use: replacing montane rain forest by subsistence cropping in Tanzania (based on original data in Edwards, 1979).

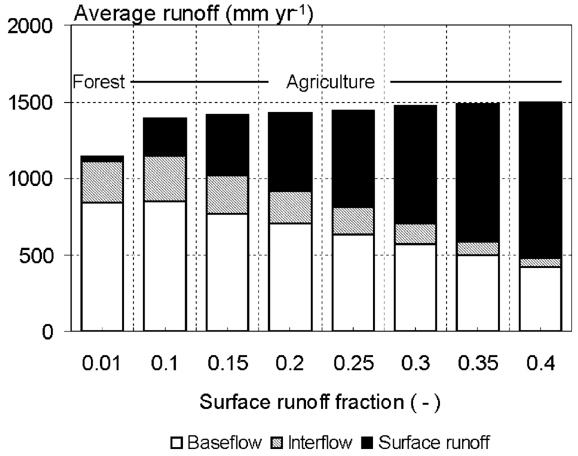

Fig. 6. Simulated changes in streamflow components for forested and agricultural conditions with increasing soil disturbance (after Van der Weert, 1994).

the continued infiltration of (most of) the rainfall, then the reduced ET associated with forest removal will show up as increased dry season flow (Fig. 5b).

Infiltration opportunities may be conserved through the establishment of a well-planned and maintained road system plus the careful extraction of timber in the case of logging operations (Bruijnzeel, 1992; Dykstra, 1996), or by applying appropriate soil conservation measures after clearing for agricultural purposes (Hudson, 1995; Young, 1989).

Although the 'infiltration trade-off' hypothesis (Bruijnzeel, 1989) remains to be proven, some support for it comes from a modelling exercise by Van der Weert (1994). The relative contributions to annual water yield by three streamflow components, viz. surface runoff ('infiltration-excess overland flow'), subsurface stormflow (called 'interflow' by Van der Weert), and baseflow (deep groundwater outflow) were simulated for fully forested and fully cleared (rainfed agricultural) conditions with gradually increasing surface runoff coefficients using a 10-year series of monthly rainfall data for the Citarum basin cited earlier (Fig. 6). Without going into detail with respect to the model used, the simulations clearly show that baseflow levels are little affected by forest clearing as long as surface runoff coefficients remain below 15% of the rainfall. Conversely, if surface runoff becomes as high as 40%, then baseflow (dry season flow) is roughly halved (Fig. 6).

Typical surface runoff coefficients associated with bench terraced rainfed agriculture on volcanic soils in upland West Java range from 16 to 18% for terraces on moderately steep slopes to 27–33% on steep slopes, with the bulk of the runoff being supplied by the relatively compact toedrains running at the foot of the terrace risers (Purwanto and Bruijnzeel, 1998; Van Dijk, 2002). Runoff coefficients for settlements in the same area varied between 38 and 68% (Purwanto, 1999). Very similar values have been reported for various road surface types in comparable terrain in East Java (Rijsdijk and Bruijnzeel, 1990, 1991). Such findings and those of Binn-Ithnin (1988) in the Kuala Lumpur area lend further support to the interpretation of the changes in water yield for the Citarum basin offered earlier.

Given the enormous differences in groundwater reserves, and therefore baseflow discharges, that may exist between catchments of contrasting size (e.g.

Hardjono, 1980) and geological make-up (e.g. deep deposits of volcanic tuffs versus shallow soils on impervious marls in Java; Roessel, 1939; Meijerink, 1977), the relative impact of land-cover change on low flows can be expected to differ accordingly. Experimental evidence from the tropics is lacking but another modelling exercise by Van der Weert (1994) strongly suggests that dry season flow diminishes more rapidly following severe surface disturbance in the case of deep soils with large storage capacity than in the case of more shallow soils having little capacity to store water anyway. Further work is urgently needed to separate climatic, vegetation, soil and geological factors in this respect. Naturally, it is of vital interest to know to what extent already reduced dry season flows can be restored again by improving infiltration and storage opportunities. We will come back to this point in the section on effects of reforestation.

There is a growing body of (mostly circumstantial) evidence from Latin America that cloud forest clearance for pasture or annual cropping may lead to decreased flows in the dry season (Stadtmüller and Agudelo, 1990; Brown et al., 1996; Ataroff and Rada, 2000). Ataroff and Rada (2000) reported surface runoff from undisturbed cloud forest and (lightly grazed) pasture in Venezuela to be similar (less than 2%). Under such conditions, changes in soil water content and streamflow can be expected to reflect the net effect of changes in vegetation water use and cloud stripping. However, in the case of conversion to annual cropping, there is the confounding effect of the corresponding changes in infiltration opportunities associated with soil degradation. Unfortunately, the available evidence is still inconclusive. The catchment pairs in Honduras and Guatemala for which Brown et al. (1996) derived a 50% reduction in dry season flow after conversion to vegetable cropping, were rather different in size and elevational range, and therefore in their exposure to fog and rainfall, thus rendering the results inconclusive. A more convincing, albeit non-tropical, case was provided by Ingwersen (1985) who observed a (modest) decline in summer flows after a 25% patch clearcut operation in the same catchment in the Pacific Northwest region of the US for which Harr (1982) had inferred an annual contribution by fog of ca. 880 mm (a very high value). The effect disappeared after 5–6 years. Because forest cutting in the Pacific Northwest is normally associated with strong increases in water yield (Harr, 1983), this anomalous result was ascribed to an initial loss of fog stripping upon timber harvesting, followed by a gradual recovery during regrowth. Interestingly, the effect was less pronounced in an adjacent (but more sheltered) catchment and it could not be excluded that some of the condensation not realised in the more exposed catchment was 'passed on' to the other catchment (Ingwersen, 1985; cf. Fallas, 1996).

4.2. Stormflows and floods: local effects

The hydrological response of small catchment areas to rainfall (stormflow production) depends on the interplay between climatic, geological and land use variables. Key parameters in this respect include the hydraulic conductivity of the soil at different depths, rainfall intensity and duration, and slope morphology (Dunne, 1978). Generally speaking, infiltration capacities of undisturbed forest soils are such that they easily accommodate most rainfall intensities (see Bonell, 1993 for a discussion of a few exceptional cases). Under such conditions, a catchment's response to rainfall usually represents a mixture of contributions by the so-called 'saturation' overland flow from wet valley bottoms and other depressions, plus rapid subsurface flow through 'macro-pores' and soil 'pipes' from the sideslopes. The relative magnitude of the respective components will vary, both between catchments as a result of differences in topography and soils (possibly also climate/rainfall intensity), and between events as a result of differences in antecedent soil moisture status and storm characteristics (Dunne, 1978). The variation in runoff response that may occur as a result of differences in soils is illustrated by the very different stormflow volumes produced by 10 very small (1 ha) undisturbed rain forest catchments in French Guyana (Fig. 7). These were all situated close to each other and therefore exposed to the same rainfall (Fritsch, 1993). Expressed as a percentage of incident rainfall, values ranged between 7.3% (catchment C) and 34.4% (catchment H). As shown in Fig. 7, a distinction can be made between basins where the groundwater table tends to be close to the surface in the valley bottom (catchments F–H), and catchments where this is not the case (other basins). In the former, storm runoff was dominated by saturation overland flow generated in the wet valley bottoms versus rapid

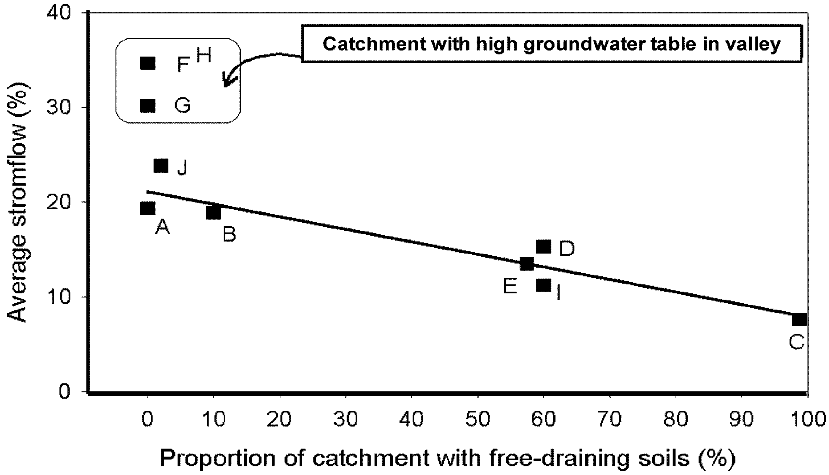

Fig. 7. Stormflow as a percentage of annual rainfall for ten nearly adjacent small catchments in French Guyana as a function of the proportion of catchment area underlain by free-draining soils (modified from Fritsch, 1993).

subsurface flow in the other areas. Interestingly, the average response of the latter group proved to be negatively related to the percentage of area underlain by well-drained soils (Fig. 7). In other words, the better the soils were drained, the smaller the runoff response (Fritsch, 1993).

Carefully planned and conducted conversion operations will be able to keep soil compaction and disturbance, and hence occurrence of infiltration-excess overland flow, to a minimum, particularly when refraining from burning (Hsia, 1987; Malmer, 1996; Swindel et al., 1982). However, even with minimum soil disturbance, there will still be increases in peakflows after forest removal, because the associated reduction in ET will cause the soil to be wetter (cf. Fig. 3) and therefore more responsive to rainfall. Relative increases in response tend to be largest for small rainfall events (roughly 100–300%), but decline to 10% or less for large events (Gilmour, 1977; Pearce et al., 1980; Hewlett and Doss, 1984). As such, the effect decreases with increasing rainfall, suggesting that soil factors begin to override vegetation factors as the soils grow wetter (see also below). Although such increases can be expected to diminish within 1–2 years as a new cover establishes itself (Hsia, 1987; Fritsch, 1993; Malmer, 1996; cf. Lal, 1996), they may become 'structural' because of contributions from roads and residential areas where there were none before (Binn-Ithnin, 1988), or because soils remain wetter throughout (e.g. in the case of a conversion to pasture; Fritsch, 1993).

Normally, peaks (and to a lesser extent stormflow volumes) produced by some form of overland flow are more pronounced than those generated by subsurface types of flow (Dunne, 1978). Therefore, the dramatic increases in peakflows/stormflows that are often reported after logging or land clearing operations using heavy machinery (Fritsch, 1992; Malmer, 1993) primarily reflect a shift from subsurface flow to overland flow dominated stormflow patterns as a result of increased soil compaction (Kamaruzaman, 1991; Van der Plas and Bruijnzeel, 1993). However, in catchments where overland flow (usually of the 'saturation' type) is already rampant under undisturbed conditions (for instance due to the presence of an impeding layer at shallow depth; Bonell and Gilmour, 1978), the response to rainfall after forest removal hardly increases any further (Gilmour, 1977). An example of soils with such poor hydraulic conductivity, and thus high surface runoff even under forested conditions in southeast Asia are the shallow heavy clay soils developed from marls in Java that are usually planted to teak. Storm runoff coefficients under such conditions are typically >30% of incident rainfall (Coster, 1938; Van Dijk and Ehrencron, 1949). Conversely, values for forested upland catchments on deep porous volcanic deposits in Java (no overland flow) are typically less than 5%, but these may increase to 10–35% after

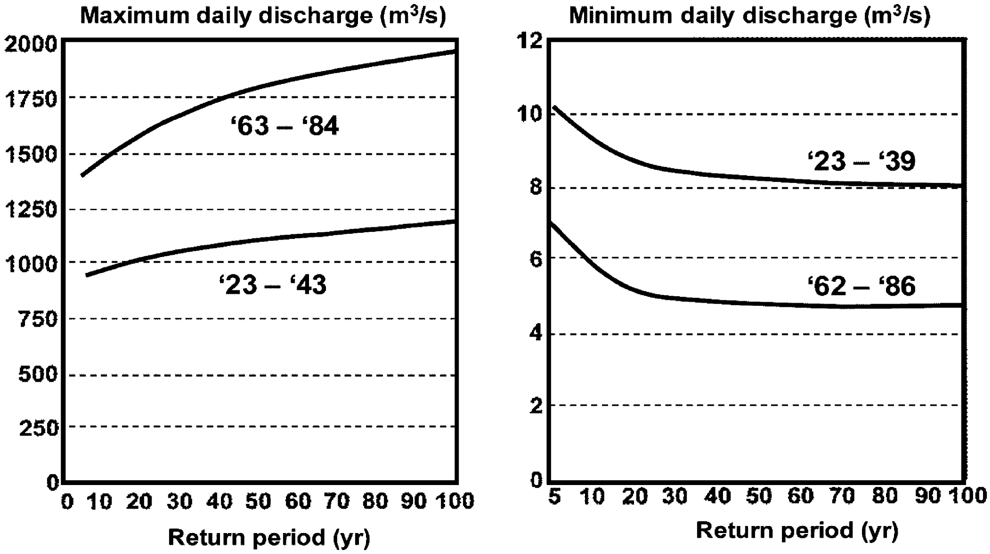

Fig. 8. Changes in average maximum and minimum daily flows for the Citarum river basin, West Java, Indonesia, between 1923–1939 and 1962/1963–1984/1986 (after Van der Weert, 1994).

clearing, depending on the proportion of the catchment occupied by settlements and roads (Rijsdijk and Bruijnzeel, 1990; Sinukaban and Pawitan, 1998; Purwanto, 1999). Binn-Ithnin (1988) reported an average increase in stormflow volumes of 250% following urbanisation in Kuala Lumpur, Malaysia, compared to forested conditions whereas peak flows were increased by more than four times (cf. Fig. 8).

4.3. Stormflows and floods: off-site effects

Whilst it is beyond doubt that adverse land use practices after forest clearance cause serious increases in stormflow volumes and peakflows, one has to be careful to extrapolate such local effects to larger areas. High stormflows generated by heavy rain on a misused part of a river basin may be 'diluted' by more modest flows from other parts receiving less or no rainfall at the time, or having regenerating vegetation c.q. better land use practices (Hewlett, 1982; Qian, 1983; Dyhr-Nielsen, 1986). Also, it is important to not immediately attribute short-term trends in the frequency of occurrence of peak discharges or floods on large river systems to upstream changes in land use. Gentry and Lopez-Parodi (1980), for example, examined the 1962–1978 water level records for the Amazon River at Iquitos, Perú, and found a distinct increase in the height of the annual flood crest that they ascribed to large-scale deforestation in the Andes. A subsequent analysis by Richey et al. (1989) of a much longer time series (1903–1983) for a gauging station further downstream (Manaus) revealed that the increase in flow during the 1962–1978 period was within the range of longer-term cycles which, in turn, were mainly determined by large-scale fluctuations in climate. Nevertheless, the recent finding of significantly increased (by 28%) wet season flows (not floods), of which the peak also arrived one month earlier, after 19% of the 175,360 km^2 Tocantins basin in Brazil had been converted to pasture (Costa et al., 2003) does illustrate the possibility of increased stormflows even at this scale. As mentioned earlier, strongly reduced infiltration opportunities under grazing were considered the main reason for this finding.

Increased wet season flows are one thing but truly devastating and large-scale floods quite another. The latter are generally the result of an equally large and persistent field of extreme rainfall (Raghavendra, 1982; Mooley and Parthasarathy, 1983), particularly when this occurs at the end of a rainy season when soils have already become wetted thoroughly by antecedent rains. Under such extreme conditions, basin response will be governed almost entirely by soil water storage opportunities rather than topsoil infiltration capacity or vegetation cover (Hewlett, 1982; Hamilton, 1987a). In other words, even in places where vegetation and soil have remained intact, the normally moderating effect of a well-developed forest cover and litter layer tends to disappear. Arguably, this does not so much impair the usefulness of the 'sponge' concept (cf. Forsyth, 1996; Calder, 2002) but rather illustrates the range of conditions under which it can be usefully applied.

Nevertheless, it cannot be excluded that *widespread* forest removal, followed by poor cultivation practices *and rampant soil degradation*, may have a cumulative effect. Indeed, there are signs (e.g. the widening of river beds) that the latter may indeed be happening in various parts of the (outer) tropics, such as in Nigeria (Odemerho, 1984) and the Himalaya (Pereira, 1989). Similar reports of deteriorating flow regimes are available for large basins (>10,000 km^2) in southern China (Chen, 1987) and Brazil (Costa et al., 2003). However, other factors, which often tend to be overlooked, include torrential rains on the flood (!) plains themselves during times of high water; backwater effects where two large rivers meet; raised river beds due to high (even natural) sedimentation rates; and last, but not least, increased urbanisation (Binn-Ithnin, 1988; Bruijnzeel and Bremmer, 1989; Zhang, 1990; Van der Weert, 1994). An example of the potentially important impact of the latter is given in Fig. 8. Despite allegedly minor changes in forest cover and peak rainfalls for the two observation periods under consideration, maximum flows in the 4133 km^2 Citarum basin, West Java, have clearly increased considerably in the latter period (Van der Weert, 1994). Finally, Hamilton (1987a) made the important point that equating economic losses associated with a flood with the severity of that flood may give an impression of more frequent and damaging events while in reality the former may rather reflect economic growth and increased floodplain occupancy.

5. Hydrological effects of (re)forestation

In response to the widely observed degradation of formerly forested land and the rising demands for paper pulp, industrial wood and fuelwood, the need for large-scale reforestation programmes has been expressed repeatedly (e.g. FAO, 1986b; Postel and Heise, 1988; Valdiya and Bartarya, 1989; cf. Brown et al., 1997). It is of great interest, therefore, to examine to what extent plantations and other conservation measures aimed at promoting infiltration are indeed capable of restoring the original hydrological conditions, i.e. not only reduce peak flows but, above all, enhance low flows as well. Evidence of reductions in peak and stormflows after reforestation and the digging of contour trenches (by 60–75%; see Bruijnzeel and Bremmer, 1989 for a review) comes from a series of paired catchment experiments in the seriously degraded Lesser Himalaya in northern India. However, streamflow from these small catchments was not perennial and therefore any effects on low flows could not be evaluated.

Despite the widespread intuitive feeling that reforestation or conservation measures will restore the flow from dried-up springs (Spears, 1982; Mann, 1989; Valdiya and Bartarya, 1989), the only apparently successful case known to the present author is that of the Sikka catchment in Flores, eastern Indonesia, as mentioned in passing by Carson (1989) and Nooteboom (1987). Because further information on the particulars of the situation (including the tree species used and the underlying geology) is lacking, it is difficult to ascertain whether this increase in flow is due to temporarily higher rainfall or indeed to a major increase in infiltration (cf. Pattanayak, this volume). Additional observations in the area may shed more light on this intriguing case. Other claims of increased dry season flows after reforestation or soil conservation in the tropical literature may be reduced to differences in catchment size and deep leakage (Hardjono, 1980), rainfall patterns between years (Sinukaban and Pawitan, 1998) or calculation errors (Negi et al., 1998).

However, there is one particular situation in which reforestation of degraded grass- or crop land may result in enhanced low flows. As indicated earlier, contributions by intercepted cloud water to the water budget of montane cloud forests may attain substantial values during rainless periods (Bruijnzeel and Proctor, 1995; Bruijnzeel, 2002a). Although the cloud stripping capacity of the original forest is largely lost upon complete conversion to pasture or short crops, it can be recreated by reforestation. Also, certain planting configurations (e.g. hedgerows or small blocks of trees positioned perpendicularly to the direction of the main air flow) help to maximise the exposure of the trees to passing fog (Kashiyama, 1956; Ekern, 1964).

Fallas (1996) demonstrated how remnant patches of cloud forest surrounded by pasture in Costa Rica intercepted at least as much cloud water as the original closed canopy forest (cf. the example from the Pacific Northwest cited earlier; Ingwersen, 1985). Again, the effect at the catchment scale can be expected to depend on the relative area occupied by such plantings or secondary growth as well as their orientation with respect to the prevailing winds.

Although there is little doubt that annual water yields from forested areas are reduced compared to those for non-forested areas (Figs. 2, 4 and 6), it must be granted that no catchment forestation experiments have investigated effects on dry season flows on *seriously degraded land*. Work to this end is in progress, however, in the state of Karnataka, India (B.K. Purandara, personal communication).

As such, it could be argued that the clear reductions in total and dry season flows observed after afforesting natural grass- and scrubland with pines or eucalypts in South Africa (Dye, 1996; Scott and Smith, 1997), South India (Samraj et al., 1988; Sharda et al., 1988; Sikka et al., 2003) and Fiji (Waterloo et al., 1999) only serve to demonstrate the difference in water use between forest and grassland. In other words, the potentially beneficial effects on low flows afforded by improved infiltration and soil water retention capacities during forest development could not become manifest in these examples. The key question is, therefore, whether the reductions in storm runoff generating overland flow incurred by such soil physical improvements can be sufficiently large to compensate the extra water use by the new forest, and so (theoretically) boost low flows (Bruijnzeel, 1989; cf. Fig. 6).

There is no easy answer to this question for several reasons. First of all, the effect of an increase in topsoil infiltration capacity on the frequency of occurrence of infiltration-excess overland flow depends equally on prevailing rainfall intensities. For instance, rainfall intensities in the Middle Hills of Nepal were generally so low that the 140 mm h^{-1} increase in infiltration capacity 12 years after reforesting a degraded pasture site with *Pinus roxburghii* made virtually no difference in the frequency of generation of overland flow (Gilmour et al., 1987). Similarly, whilst infiltration capacities had roughly doubled in 12 years since reforesting *Imperata* grassland with teak in Sri Lanka, at 30 mm h^{-1} the hydrological impact of this increase must be very limited (Mapa, 1995). Secondly, the reductions in catchment response to rainfall observed after forestation will also reflect the drier soil conditions prevailing under actively growing forest rather than a reduction in peak-generating overland flow per se (cf. Hsia, 1987). However, with the exception of Lal (1997), none of the studies documenting decreases or increases in stormflows after land use change have attempted to quantify the associated changes in relative contributions by subsurface- and overland flow types (see reviews by Bruijnzeel, 1990; Bruijnzeel and Bremmer, 1989). Thirdly, there is also the effect of soil depth which determines both the maximum amount of water that can be stored in a catchment under optimum infiltration conditions, and the possibilities for water uptake by the developing root network of the new trees (Trimble et al., 1963). Naturally, where soils are intrinsically shallow for geological reasons (e.g. on very steep mountain slopes or on impermeable substrates subject to intense natural erosion and mass wasting; Coster, 1938; Meijerink, 1977; Ramsay, 1987a,b), or where soils have become shallow due to continued intense erosion after clearing, soil water storage opportunities are decreased accordingly (Bruijnzeel, 1989). And last, but not least, there is the confounding effect of rainfall patterns (e.g. seasonal versus well distributed) and general evaporative demand of the atmosphere, which both exert a strong influence on tree water use, particularly in sub-humid areas (Sandström, 1998). An example of the latter concerns the ‘spring sanctuary development project’ in northern India where attempts are being made to revive the flow from a nearly extinct spring by a combination of tree planting, exclusion of grazing and the digging of contour trenches. Unfortunately, the results of this interesting experiment were confounded by rainfall variability (Negi et al., 1998). Needless to say, differences in soil and climatic factors, plus the absence of detailed information on prevailing stormflow mechanisms before and after forestation, all render comparisons between different sites more complicated. Rigorous experimental designs are needed (see also Section 8).

The only documented ‘real world’ case in which the infiltration compensation mechanism seems to have occurred may be the White Hollow catchment in Tennessee, US (Tennessee Valley Authority, 1961). Prior to improvement of its vegetation cover, two-thirds of this catchment consisted of mixed secondary forest in poor condition (due to fire, heavy logging and grazing), with another 26% under poor scrub. About 40% of the catchment was estimated to be subject to more or less severe erosion at the start of the remedial measures. Following extensive physical and vegetative restoration works, peak discharges decreased considerably within 2 years, especially in summer. However, neither annual water yield nor low flows changed significantly over the next 22 years of forest recovery and reforestation. It was concluded that the extra water needed by the recovering and additionally planted trees was balanced by improved infiltration (Tennessee Valley Authority, 1961). It is quite possible, however, that the absence of major changes in total and low flows at White Hollow mainly reflects the lack of contrast between tall and short vegetation, as would have been the case when reforesting a truly deforested and degraded catchment. Support for this contention comes from the observation of Trimble et al. (1987) that reductions in flow from several large river basins in the southeastern US, which had suffered considerable erosion before being partially reforested, were largest during dry years. In addition, the effect became more pronounced as the trees grew older. A similar case was recently described for a once seriously degraded catchment in the Mediterranean part of Slovenia (former Yugoslavia), where the spontaneous return of (deciduous) forest produced a steady reduction in annual water yield over a period of 30 years, with the strongest reductions again occurring during the dry summer months (Globevnik and Sovinc, 1998). Therefore, in both these real-world examples the water use aspect overrides the infiltration aspect, despite the rather modest contrasts in water use between forest and grass/crop land under the prevailing climatic conditions (e.g. Hibbert, 1969).

Unfortunately, such findings offer little prospect for the possibility of significantly raising dry season flows in the humid tropics by forestation with fast-growing (usually exotic) species, despite claims to this extent (e.g. Cheng, 1999; Hardjono, 1980). Observed maximum differences between annual water use of pines or eucalypts and short vegetation (grass, crops) under well-watered (sub-)tropical conditions attain values of 500–700 mm at the catchment scale (Fig. 9) and even higher values (>1000 mm) on individual plots with particularly vigorous tree growth (Dye, 1996; Waterloo et al., 1999). The hydrological benefits

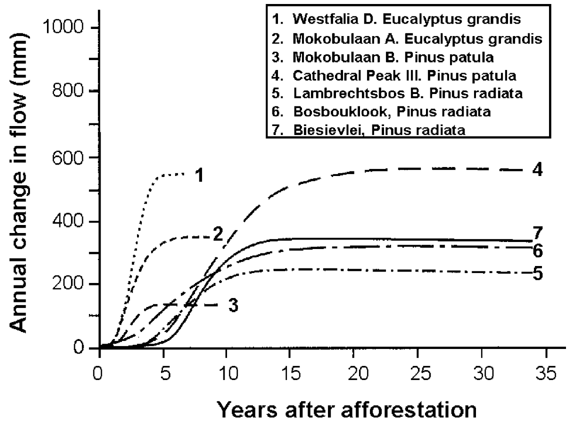

Fig. 9. Trends in post-afforestation changes in catchment water yield for seven paired catchment experiments in different forestry regions of South Africa (based on data in Bosch, 1979; Dye, 1996).

incurred by the limited increases in topsoil infiltration capacities observed after forestation of degraded land in Nepal and Sri Lanka cited earlier, are simply dwarfed by such high water requirements. To make matters even worse, the water use of the most commonly planted tree species peaks much earlier than the time periods usually required for a full recovery of soil infiltration capacity (Fig. 9; Gilmour et al., 1987). The conclusion that already diminished dry season flows in degraded tropical areas may decrease even further upon reforestation with fast-growing tree species seems inescapable, therefore (Bruijnzeel, 1997). However, sufficiently large reductions in hillslope runoff response to rainfall to potentially boost low flows were implied by the comparative measurements of Chandler and Walter (1998) for severely degraded grassland, conservation cropping (mulching) and semi-mature secondary growth (15–20 years) in the philippines. Unfortunately, their study did not include measurements of streamflow or the groundwater table. In view of the extent of the low flow problem, the testing of alternative ways of increasing water retention in tropical catchments without the excessive water use normally associated with exotic tree plantations should receive high priority. One could think in this respect of (a combination of) physical conservation measures (e.g. bench terracing with grassed risers, contour trenches, runoff collection wells in settlement areas: Roessel, 1939; Negi et al., 1998; Purwanto, 1999; Van Dijk, 2002), vegetative 'filter' strips at strategic points in the landscape (Dillaha et al., 1989; Van Noordwijk et al., 1998), and the use of indigenous species with perhaps lower water use (Negi et al., 1998), possibly in an agroforestry context in which the trees may also assist in maintaining slope stability (Young, 1989; O'Loughlin, 1984). Calder (1999) suggested rotational land use, in which periods with forest alternate with periods of agricultural cropping, as a potential way of reducing long-term mining of soil water reserves by the trees. Naturally, for this approach to be successful at all, soil degradation during the cropping phases should be avoided.

6. Tropical forests and catchment sediment yield

6.1. General considerations

Findings such as that overall streamflow amounts from non-forested areas are higher than those associated with forested areas (Figs. 2 and 9), or that changes in stormflow volumes ('floods') are determined less by the presence or absence of a forest cover as rainfall becomes more extreme, should not be taken to imply that forest removal could not have serious adverse consequences (Smiet, 1987). On the contrary, surface erosion and catchment sediment yield normally show dramatic increases in such cases (Gilmour, 1977; Fritsch, 1992; Douglas, 1996; Fig. 10). When dealing with the effects of changes in land use on erosion and sedimentation, it is helpful to distinguish between surface erosion, gully erosion, and mass movements, because the ability of a vegetation cover to control these various forms of erosion is rather different. Sediment production under natural, forested conditions may be vastly different, depending on the relative importance of the respective contributing mechanisms (Pearce, 1986). For example, suspended sediment yields from rain forested catchments may be as low as 0.25 t ha^{-1} per year in tectonically stable areas underlain by soils that are neither subject to significant surface erosion nor extensive gullying or mass wasting (Douglas, 1967; Malmer, 1990; Fritsch, 1992; Fig. 10, category I). Conversely, in tectonically active steepland areas prone to hillslope failure, as in the Himalaya and along the Pacific rim, this may increase to 35–40 t ha^{-1} per year (Pain and Bowler, 1973; Li, 1976; Ramsay, 1987a,b; Dickinson et al., 1990), whereas still higher values have been reported for situations where both surface erosion and mass wastage are rampant, such as on marls in Java (up to 65 t ha^{-1} per year; Coster, 1938; Van Dijk and Ehrencron, 1949; Fig. 10, category IV). Clearly, any effects of forest clearing on catchment sediment yield will be much more pronounced in areas where natural rates of sediment tend to be low. As such, caution is needed when comparing sediment yield figures for different regions. Also, the fact that a considerable portion of the material delivered to streams by (deep-seated) mass movements tends to be rather coarse and will be transported mainly as (generally unmeasured) bedload represents a further complication in this respect (Pickup et al., 1981; Simon and Guzman-Rios, 1990).

It is equally important to make a distinction between on-site erosion and downstream (or 'off-site') effects, because not all of the eroded material will enter the drainage network (streams) immediately. Part of it may become temporarily deposited in depressions, on footslopes or alluvial plains, etc. Therefore, effects of soil disturbance can be observed much earlier on the hillslopes themselves (in the form of increased surface erosion and loss of plant productivity) than further downstream (as increases in catchment sediment yield). Because the number of sediment storage opportunities generally increases with catchment size, the time lag between on-site events and off-site effects tends to increase with catchment size as well (Walling, 1983; Pearce, 1986). It follows that both aspects need to be taken into account if a clear understanding is to be obtained. A pertinent example includes the study of hillslope surface erosion and sediment yield patterns during the transition from a (pine) forest cover via regenerating scrub to rainfed agriculture on steep slopes in a small catchment in the volcanic uplands of West Java by Bons (1990). Catchment sediment yields were observed to increase immediately when farmers started to clear the scrub to make way for tobacco and vegetables. However, it proved premature to attribute this rise in sediment yield to the agricultural activity on the hillslopes because on-site measurements revealed that no erosion was occurring in the freshly cleared fields. Rather, the farmers had begun to remove logs that had fallen into the stream during the previous logging operation (and which had been abandoned by the foresters) for use as fuelwood. In the process, sediment was released that

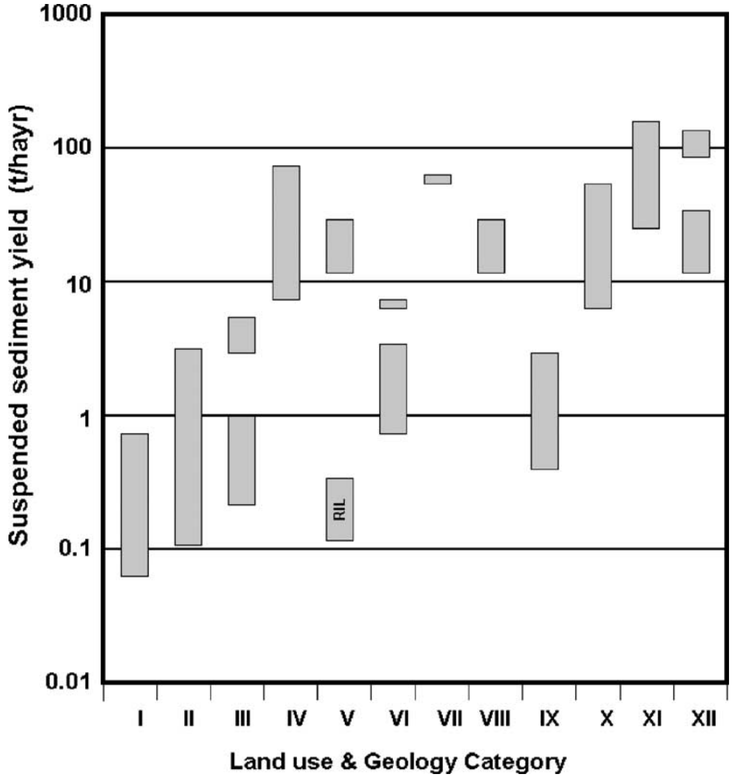

Fig. 10. Ranges in reported catchment sediment yields in southeast Asia as a function of geological substrate and land use. Categories: I, forest, granite; II, forest, sandstones/shales; III, forest, volcanics; IV, forest, marls; V, logged (RIL: reduced impact logging); VI, cleared, sedimentary rocks (lower bar: micro-catchments); VII, cleared, volcanics; VIII, cleared, marls; IX, medium-large basins, mixed land use, granite; X, idem, volcanics; XI, idem, volcanics plus marls; XII, urbanised (lower bar), mining and road building (upper bar).

had accumulated previously behind the log dams when the stream banks became damaged during logging (Bons, 1990). Similarly, Fritsch and Sarrailh (1986) explained a dramatic contrast in stream sediment load ($6\,t\,ha^{-1}$ per year) and sideslope soil displacement by heavy machinery (equivalent to $1200\,t\,ha^{-1}$ per year) during a forest clearing operation in French Guyana by the filtering effect of a wall of earth and logging debris that had accumulated around a valley bottom that was too wet to permit access to machinery at the time. A third point relates to the fact that stream sediment loads tend to vary enormously in time, with values being disproportionately higher during very wet periods or years, or even during individual extreme events (Walling, 1983; Dickinson et al., 1990; Douglas et al., 1999). As such, it is imperative that peak events are sampled properly or else sediment yields will be seriously underestimated. A pertinent example was given by Biksham and Subramanian (1988) for the Godavari river in Central India. Based on occasional sampling during the respective seasons over a period of 3 years, the annual suspended sediment load was underestimated by some 240% compared to the estimate based on daily sampling. Also, the effects of truly extreme events that suddenly deliver very large amounts of sediment to the drainage system (such as earthquakes, hurricanes or volcanic eruptions; Goswami, 1985; White, 1990; Douglas, 1996), may be visible long after the event itself, especially on very large river systems. An extreme example is provided by the Brahmaputra River where the river bed continued to rise over a stretch of 600 km for 21 years after a major earthquake occurred in August 1950. Once sediment inputs to the river decreased in the early 1970s the river slowly started to degrade its bed again, thereby increasing overall sediment concentrations at downstream stations (Goswami, 1985). It follows that caution is needed when interpreting changes in sediment yield over time at a particular gauging station and that such changes may not always be directly attributable to 'deforestation' (e.g. Brabben, 1979; Narayana, 1987).

With the above caveats in mind, what does research have to offer with respect to the influence of forest removal and subsequent land management on sediment production (erosion) and catchment sediment yield in the humid tropics?

6.2. Surface erosion

This form of erosion is rarely significant in areas where the soil surface is protected against the direct impact of the rain, be it through a litter layer maintained by some sort of vegetation or through the application of a mulching layer in an agricultural context. The results of about 80 studies of surface erosion rates in tropical forest and tree crop systems have been collated in Table 1 (after Wiersum, 1984). Although the data are of variable quality and pertain to a variety of soil types, they clearly show surface erosion to be minimal in those cases where the soil is adequately protected (categories 1–4). Erosion rates increase somewhat upon removal of the understorey (category 5) but rise dramatically only when the litter layer is removed or destroyed (categories 8–9). The initial effect is rather small due to the effect of residual organic matter on soil aggregate stability and infiltration capacity (nos. 6 and 7; Gonggrijp, 1941a; Wiersum, 1985; Lal, 1996) but becomes considerable upon repeated disturbance of the soil by burning, frequent weeding or overgrazing, which all tend to make the soil compacted or crusted, with impaired infiltration and accelerated erosion as a result (Jasmin, 1975; Costales, 1979; Toky and Ramakrishnan, 1981).

The results collated in Table 1 confirm the important point made by Smiet (1987) that the margins for forest management with respect to soil surface protection against erosion are much broader than those associated with grazing or annual cropping. Whereas the degraded natural and plantation forests of many tropical uplands are still able to fulfil a protective role because gaps are usually rapidly colonised by pioneer species (undergrowth), grasslands are often more prone to fire, overgrazing and landsliding (Jasmin, 1975; Gilmour et al., 1987; Haigh, 1984). On the other hand, erosion on well-kept grassland (Fritsch and Sarrailh, 1986), moderately grazed forest (Wiersum, 1984) and agricultural fields with appropriate soil conservation measures on otherwise stable slopes (Hudson, 1995; Paningbatan et al., 1995; Young, 1989) is usually low.

There is increasing evidence that erosion rates on and around such compacted surfaces as skidder tracks and log landings, roads, footpaths and settlements can be very high, especially shortly after construction (35–500 t ha^{-1} per year; Henderson and Witthawatchutikul, 1984; Bons, 1990; Rijsdijk and Bruijnzeel, 1990, 1991; Nussbaum et al., 1995; Malmer, 1996; Purwanto, 1999). In addition, the very considerable volumes of runoff generated by such surfaces may promote downslope gully formation and mass wastage. Therefore, as already noted for runoff (Fig. 8), sediment contributions to the stream network by roads and settlements may be disproportionately large for their relatively small surface area (see also below). More work is needed to incorporate such

Table 1
Surface erosion rates in tropical forest and tree crop systems (t ha^{-1} per year; after Wiersum, 1984)

	Minimum	Median	Maximum
1. Natural forests (18/27)[a]	0.03	0.3	6.2
2. Shifting cultivation, fallow phase (6/14)	0.05	0.2	7.4
3. Plantations (14/20)	0.02	0.6	6.2
4. Tree gardens (4/4)	0.01	0.1	0.2
5. Tree crops with cover crop/mulch (9/17)	0.10	0.8	5.6
6. Shifting cultivation, cropping phase (7/22)	0.4	2.8	70
7. Agricultural intercropping in young forest plantations ('taungya') (2/6)	0.6	5.2	17.4
8. Tree crops, clean-weeded (10/17)	1.2	48	183
9. Forest plantations, litter removed or burned (7/7)	5.9	53	105

[a] (a/b) = number of locations/number of 'treatments'.

contributions in the current generation of erosion and catchment sediment yield models (cf. De Roo, 1993; Ziegler et al., this volume).

6.3. Gully erosion

Gully erosion is a relatively rare phenomenon in most rain forests but may be triggered during extreme rainfall when the soil becomes exposed through treefall or landslips (Ruxton, 1967). In other cases, gullies may form by the collapse of subsurface soil 'pipes' (Morgan, 1995). As indicated earlier, active gullying in formerly forested areas is often related to compaction of the soil by overgrazing or the improper discharging of runoff from roads, trails and settlements (Bergsma, 1977). Poesen et al. (2003) stressed the importance of gullies to catchment sediment yield in view of the increased 'connectivity' afforded by gullies between hillslope fields and streams. If gullies are not treated at an early stage, they may reach a point where restoration becomes difficult and expensive. The moderating effect of vegetation on actively eroding gullies is limited and additional mechanical measures such as check dams, retaining walls and diversion ditches will be needed (Blaisdell, 1981; FAO, 1985, 1986a).

6.4. Mass wasting

Mass wasting in the form of deep-seated (>3 m) landslides is not influenced appreciably by the presence or absence of a well-developed forest cover. Geological (degree of fracturing, seismicity), topographical (slope steepness and shape) and climatic factors (notably rainfall) are the dominant controls (Ramsay, 1987a,b). However, the presence of a forest cover is generally considered important in the prevention of shallow (<1 m) slides, the chief factor being mechanical reinforcement of the soil by the tree root network (Starkel, 1972; O'Loughlin, 1984). Bruijnzeel and Bremmer (1989) cite unpublished observations by I.R. Manandhar and N.R. Khanal on the occurrence of shallow landslides in an area underlain by limestones and phyllites in the Middle Hills of Nepal. Most of the 650 slips that were recorded between 1972 and 1986 had been triggered on steep (>33°) deforested slopes during a single cloudburst whereas only a few landslides had occurred in the thickly vegetated headwater area. However, under certain extreme conditions, such as the passage of a hurricane, the presence of a tall tree cover may become a liability in that trees at exposed locations may be particularly prone to becoming uprooted, whereas, in addition, the weight of the trees may become a decisive factor once the soils reach saturation. Scatena and Larsen (1991) reported that out of 285 landslides associated with the passage of hurricane Hugo over eastern Puerto Rico, 77% occurred on forest-covered slopes and ridges. More than half of these mostly shallow landslips were on concave slopes that had received at least 200 mm of rain. Brunsden et al. (1981) described a similar case in eastern Nepal where mass wasting on steep forested slopes was much more intensive than in more gently sloping cultivated areas. Although often occurring in large numbers, such small and shallow slope failures usually become quickly revegetated and, because of their predominant occurrence on the higher and central portions of the slopes, contribute relatively little to overall stream sediment loads, in contrast to their more deep-seated counterparts (Ramsay, 1987a).

6.5. Catchment sediment yields

It will be clear from the foregoing that no 'typical' values can be given for the changes in catchment sediment yield upon tropical forest disturbance or conversion. Nevertheless, a fairly consistent picture emerges from Fig. 10 which summarises the results obtained by more than 60 studies of (suspended) sediment yield from small to medium-sized (generally $\ll$100 km^2) catchments in southeast Asia as a function of geological substrate, land cover and degree of disturbance (based on the following sources for Malaysia: Lai, 1993; Baharuddin and Abdul Rahim, 1994; Greer et al., 1995; Douglas, 1996; for Indonesia: Van der Linden, 1978, 1979; Bons, 1990; Rijsdijk and Bruijnzeel, 1990, 1991; Purwanto, 1999; for the Philippines: Dickinson et al., 1990; and for Thailand: Alford, 1992).

Under undisturbed forested conditions, suspended sediment yields are generally below 1 t ha^{-1} per year for very small (<50 ha) headwater catchments, regardless whether these are underlain by granitic, young volcanic or sedimentary rocks (Fig. 10, categories I–III). Somewhat higher values (typically 3–5 t ha^{-1} per year) are obtained for forested catchments of a few square kilometres in size on sedimentary rocks and young volcanics, whereas a much higher sediment yield (66 t ha^{-1} per year) was reported for a forested catchment of intermediate size (45 km^2) in Central Java on unstable marly soils (Van Dijk and Ehrencron, 1949; Fig. 10, category IV). Because pre-war observations in Java were based on spot measurements at fixed times during the day, and because of the highly peaked nature of the runoff from marly areas, the quoted figure must be considered an underestimate, possibly by as much as a factor 2 (D.C. van Enk, personal communication).

The construction of roads, skidder tracks and log landings during mechanised logging and clearing operations represents a serious disturbance to the forest and generally causes sediment yields to rise 10–20 times (Fig. 10, categories V and VI, respectively). However, the effect usually subsides within a few years as skidder tracks become revegetated, roadsides stabilise and (in the case of clearing) the new vegetation establishes itself (Douglas et al., 1992; Malmer, 1996), although the stored sediment may be remobilised during extreme events even after many years (Douglas et al., 1999). Increases in sediment yields associated with reduced impact logging (RIL) are generally much lower than for the average commercial operation (Baharuddin and Abdul Rahim, 1994; Fig. 10, category V). The low impact of forest clearing in very small catchments in East Malaysia (Fig. 10, lower part of category VI) probably relates to the trapping of eroded material by logging debris because skidder tracks in the area were observed to erode at the very high rate of ca. 500 t ha^{-1} per year (Malmer, 1996). High sediment yields have also been recorded for small agricultural upland catchments in young volcanic (up to 55 t ha^{-1} per year; Fig. 10, category VII) and marly (10–27 t ha^{-1} per year; category VIII) terrain in Java. Sediment yields of medium-sized catchments with mixed land use (including forest in various stages of regrowth) increase in the sequence: granite $<$ young volcanics $<$ marls (Fig. 10, categories IX, X and XI, respectively). Finally, the dramatic effects of such drastic disturbances as urbanisation, mining and road building are clearly borne out by the few data that are available, regardless of geological substrate (Fig. 10, category XII; Douglas, 1996; Pickup et al., 1981; Henderson and Witthawatchutikul, 1984).

Overall sediment yield figures for medium-sized catchments harbouring a variety of land-cover types do not give information on the main source(s) of the sediment. Yet it is obviously of great practical importance for the design (and evaluation) of soil conservation schemes to know what portions of the catchment, or what processes, contribute most of the sediment. Once again, the physiographic setting is of paramount importance. For example, Fleming (1988) calculated that a reduction in soil loss from heavily overgrazed lands in the Phewa Tal catchment in the Middle Mountains of Nepal to a level representative of improved pastures would have a negligible effect (ca. 1%) on the rate at which a downstream lake was silting up because 95% of the sediment entering the lake was derived from geologically controlled deep-seated mass wasting and bank erosion (Ramsay, 1987a). A rather different picture was obtained for the equally steep volcanic Konto catchment (233 km^2) in East Java, Indonesia, which had ca. two-thirds of its area under (degraded) forest, the remainder being used for intensive agriculture (both rainfed and irrigated) and settlement areas. Mass wasting and bank erosion were estimated to contribute only 9% to the total sediment yield, whereas roads, settlements and trails (together occupying ca. 5% of the total area) contributed roughly 54%, with the remaining 37% of the sediment coming from the ca. 20% of the area occupied by rainfed agriculture (recomputed from data in Rijsdijk and Bruijnzeel, 1991). As such, there would seem to be considerable scope for reducing sediment production in this particular area. In view of the disproportionately large influence on runoff and sediment generation exerted by roads, trails and settlements, particular attention would need to be paid to the proper discharging c.q. trapping of excess runoff and sediment from such areas. Purwanto (1999) related in this respect how infiltration wells were installed in upland villages in West Java to not only intercept runoff but also boost infiltration and baseflows. Unfortunately, the openings of the wells tended to become rapidly blocked by garbage carried by the runoff, thereby seriously reducing their efficiency (L.A. Bruijnzeel, personal observation).

Although there are certain geological controls on catchment sediment yield (notably the presence or absence of unstable marly soils), an important conclusion that may be drawn from Fig. 10 is that increases in sediment yields during logging and clearing operations can be kept low by reduced impact logging techniques which minimise surface disturbance (Malmer, 1990, 1996; Baharuddin and Abdul Rahim, 1994; cf. Bruijnzeel, 1992; Dykstra, 1996). Likewise, although published data on the positive effect on sediment yield incurred by soil conservation measures or reforestation in the humid tropics seem to be limited to zero-order catchments (i.e. having no perennial flow; e.g. Gonggrijp, 1941a; Amphlett, 1986; Amphlett and Dickinson, 1989; Bruijnzeel and Bremmer, 1989; DaZo, 1990), an equally large effect may be expected in catchments not plagued by extensive mass wasting, such as the Konto area in East Java cited earlier. Further work is needed to check this contention over a range of scales. It is important to realise in this respect that, whatever the kind of measures taken to reduce on-site sediment production, their effect tends to become less recognisable as one proceeds further downstream. A related, and frequently overlooked, aspect is the time scale at which any downstream benefits from upland rehabilitation activities are likely to become noticeable (Pearce, 1986). In large river basins, there may be so much sediment stored in the drainage system itself that it effectively forms a long-term supply, even if all human-induced sediment inputs in the headwater area would be eliminated (cf. the example of the Brahmaputra given earlier; Goswami, 1985). Results from a major land rehabilitation project in China suggest that reductions in sediment yield of up to 30% may be expected after about 20 years for very large catchments (100,000 km^2). A significant part of this reduction was effected by the trapping of sediment behind numerous check dams and sand traps (Mou, 1986). As such, the frequently voiced expectation that upland reforestation will solve downstream problems does require some specification of the spatial and time scales involved. It is important not to raise unrealistic expectations in this respect (Goswami, 1985; Pearce, 1986). Highland–lowland interactions in tropical river basins are understudied and further work is needed.

7. Research needs

Having reviewed the various hydrological impacts of tropical forest conversion in the preceding sections, what are the chief remaining gaps in knowledge hindering the sound management of soil and water resources? Answering this question is perhaps not as straightforward as it would seem at first sight as answers coming from different interest groups are likely to differ (cf. Tomich et al., this volume). There are those who believe that 'greater emphasis on biophysical research strategies (including hillslope hydrology) may hold the key to (solving) land management problems in the humid tropics' (Bonell and Balek, 1993). Others (Hamilton and King, 1983; Pereira, 1989; Bruijnzeel, 1986, 1990) emphasised the need to 'put into practice what is known already'. The answer lies probably somewhere in the middle. Below, a selection (modified and updated from Bruijnzeel, 1996) is presented of what this author perceives as the most pressing gaps in our understanding of the hydrological consequences of land-cover transformations in the humid tropics in general, and in southeast Asia in particular.

7.1. *Effects of forest conversion on regional rainfall patterns*

In the light of the continuously rising demands for water in the region (Rosegrant et al., 1997; Abdul Rahim and Zulkifli, 1999) the trends towards increased aridity observed in different parts of south and southeast Asia (Sri Lanka: Madduma Bandara and Kuruppuarachchi, 1988; Java: Wasser and Harger, 1992; possibly northern India: Valdiya and Bartarya, 1989) are a matter of potentially great concern. There is scope for a rigorous analysis of long-term rainfall records for the respective regions to ascertain the persistence and spatial extent of such trends. Such an analysis would not only have to take into account the various cyclic fluctuations referred to earlier, but also examine whether decreases in rainfall at lowland stations are perhaps paralleled by increases higher up in the mountains (cf. Fleming, 1986). Upland rainfall stations may be identified in areas that have remained under forest throughout the observation period and others which have experienced large-scale forest removal. The analysis could be modelled after the successful FRIEND-AOC program which recently analysed climatic variability in humid Africa along the Gulf of Guinea (Servat et al., 1997; Paturel et al., 1997). The choice of prospective research areas should perhaps be guided not only by considerations of data availability but also by the observation of Koster et al. (2000) that the greatest potential influence of land-cover change on climate may be expected to occur in transition zones from humid to sub-humid climates.

In addition, although the radiative characteristics of secondary vegetation in mainland southeast Asia have been shown to resemble those of the original forest within ten years after clearing, the albedo of *Imperata* grassland is distinctly higher than that of forest (Giambelluca et al., 1999). As such, it would be of particular interest to examine through meso-scale climatic modelling approaches (cf. Lawton et al., 2001; Van der Molen, 2002) to what extent the widespread occurrence of such grasslands in parts of the region (e.g. in the Philippines, Quimio, 1996; South Kalimantan, MacKinnon et al., 1996) may have influenced local climate (e.g. intensity of sea breezes), as has been suggested by some (Nooteboom, 1987). The same may apply to areas that undergo rapid urbanisation such as Java. However, before meaningful results can be obtained with the climatic modelling approach, the parameterisation of the rain forests, grasslands and cities of southeast Asia needs to be improved. Previous deforestation simulations (Polcher and Laval, 1994; Henderson-Sellers et al., 1996) had to rely heavily on parameter values derived for forest and pasture in continental central Amazonia which may not be applicable under the more 'maritime' conditions prevailing in Malaysia, Indonesia and the Philippines (cf. Koster et al., 2000; Dolman et al., 2004). For example, there are indications that rainfall interception in southeast Asia may be at least two times those reported for central Amazonia, possibly because of large-scale advection from the surrounding warm seas (cf. Table 2 in Schellekens et al., 2000). Finally, with the planned reforestation of several million hectares of *Imperata* grasslands with fast-growing tree species in Kalimantan (C. Cossalter, personal communication) a unique opportunity may present itself to study the effects (if any) of large-scale tree planting on sub-regional rainfall and so verify the predictions of meso-scale atmospheric model simulations.

7.2. Effects of land-cover change on low flows

Although the trend of the change in total water yield following tropical forest clearing is well established, the observed variations are such that no real quantitative predictions can be made for any particular area (Fig. 2). As such, there is a distinct need to supplement the traditional paired catchment approach with process measurements and physically-based model applications (e.g. TOPOG; Vertessy et al., 1993, 1998). Arguably, such combined work is especially urgent with respect to the determination of the effects on low flows of: (i) the planting of (fast-growing, exotic) trees; (ii) the implementation of different soil conservation techniques (alone or in combination), including bench terracing, mulching, vegetative filter strips, runoff collection wells in settlements, contour trenches, agroforestry systems, etc.; and (iii) the conversion of montane cloud forest to agricultural cropping or grazing land. As indicated earlier, such work should be conducted preferably in the form of paired catchment studies, complemented with process-based measurements and modelling.

For the successful application of distributed hydrological process models that are capable of representing complex feedback mechanisms between climate, vegetation and soils (e.g. effects of soil depth on forest water use), much more information is needed on the hydrological characteristics of the various post-forest land-cover types, including upland crops (cf. Giambelluca et al., 1996, 1997, 1999; Bigelow, 2001; Van Dijk and Bruijnzeel, 2001a; Van Dijk, 2002). Also, much more needs to be known of the associated changes in soil hydraulic and water retention characteristics (Bonell, 1993; Elsenbeer et al., 1999; Godsey and Elsenbeer, 2002), and rooting patterns (Nepstad et al., 1994; Coster, 1932a,b). As to the former, additional measurements are needed of the water use and rainfall interception characteristics as a function of stand age for the most widely planted tree species, including: *A. mangium*, *G. arborea*, *P. falcataria*, *Grevillea robusta*, as well as (to a lesser extent) eucalypts and pines, teak and mahogany, for which an increasing body of information is already available (summarised by Bruijnzeel, 1997; Scott et al., 2004). Given the often adverse effects on streamflow of planting exotic tree species, due attention should also be given to comparative evaluations of indigenous species (Bigelow, 2001). Ideally, these measurements should cover a range of soil and climatic conditions, but this is obviously time-consuming. An effective way forward might be to initially make comparative observations at a limited number of sites, such as university campuses or forest research institutional grounds which often have block plantings of the most widely used species. There is considerable scope in this respect for the use of plant physiological approaches to tree water use, such as the heat pulse velocity and heat balance techniques (Smith and Allen, 1996). The latter would be equally useful in the evaluation of tree water use within the context of agroforestry systems (cf. Ong and Khan, 1993); when making comparisons between stands at different positions in the terrain (e.g. dry ridges versus moist footslopes; areas with poor growth due to soil compaction), or where shallow soils underlain by fractured rocks preclude the use of more traditional hydrological (water budget) approaches. In view of the important feedback exerted by soil moisture on vegetation water use in sub-humid areas, special attention will need to be paid to ascertaining rooting patterns and changes therein with vegetation age (cf. Coster, 1932a,b).

Also, the current lack of detailed quantitative information on changes in topsoil infiltration capacity, soil hydraulic conductivity profiles with depth, soil water retention and storage capacities, and rooting depths associated with land-cover change in the tropics calls for systematic sampling campaigns if physically-based catchment models are to be used profitably to predict the associated effects on streamflow. The database seems to be particularly weak for rainfed upland crops, plantations and secondary growth older than ca. 15 years. New sampling campaigns could usefully follow a 'false time series' approach (Gilmour et al., 1987; Giambelluca et al., 1999; Waterloo et al., 1999). On a more cautionary note, Bonell (1993) warned that the transfer of physically-based hydrological models to the humid tropics may not be without complications. For instance, the surface of the underlying bedrock may not be parallel to that of the topography in deeply weathered tropical terrain. In such cases the commonly made assumption that the hydraulic gradients of subsurface flow will run parallel to the soil surface (O'Loughlin, 1990) may not be valid (cf. Quinn et al., 1991). Indeed, a recent study in eastern Puerto Rico, which employed geophysical techniques to map subsurface topography, found the weathering surface (fresh rock) to be parallel to the general gradient of the main stream throughout the catchment area rather than to surface topography (Schellekens, 2000).

As for the hydrological effects of cloud forest clearance, any changes in water yield will presumably reflect the trade-off between the loss of the cloud water interception component upon replacement of the forest by a shorter vegetation and the net change in water use by the old and the new vegetation. Process-based studies are urgently needed of the water use and degree of cloud and rainwater interception along altitudinal gradients, preferably within the context of a paired catchment framework in view of the often leaky conditions in steep headwater areas (cf. Bruijnzeel, 2002b).

A study to this effect funded by the Department for International Development of the UK by the Vrije Universiteit Amsterdam and partners in the UK, Switzerland, Germany, USA and Costa Rica started in spring 2002 in the Monteverde area, Costa Rica.

7.3. Effects of land-cover change on runoff and sediment production

Most experimental evidence available to date (Bruijnzeel, 1990; Malmer, 1992; Fritsch, 1993) suggests that no major increases in stormflow volumes occur after *controlled* deforestation. Therefore, the adverse effects of forest removal can be kept to a minimum by adhering to a number of well-documented guidelines in most cases (cf. Pearce and Hamilton, 1986; Dykstra, 1996). As such, the widely observed contrast in this respect between theory and practice is more a socio-economic rather than a technical problem (Hamilton and King, 1983; Bruijnzeel, 1986; Pereira, 1989). Nevertheless, there is a lack of studies dealing with the effects of urbanisation on stormflow and more work is needed to incorporate the effects of roads and settlements in catchment hydrological models (De Roo, 1993; Ziegler et al., 2004). Furthermore, there is a dearth of information on the effects of land rehabilitation on catchment sediment yield, be it through physical soil conservation schemes, the introduction of agroforestry-based cropping systems or full-scale reforestation (Bruijnzeel, 1990, 1997). The database is particularly weak with respect to the time lag between land rehabilitation efforts and any subsequent reductions in stormflow and sediment transport at increasingly large distances downstream (cf. Pearce, 1986; Bruijnzeel and Bremmer, 1989; Van Dijk, 2002).

Effects of land use change on the magnitude of flood peaks in large rivers are difficult to evaluate because such changes are rarely fast and consistent (except perhaps where population pressure is very high) and often compounded by climatic variability

(Richey et al., 1989; cf. Elkaduwa and Sakthivadivel, 1999). Also, such analyses require long-term high quality data on land use, streamflow and rainfall, which are often not available in the humid tropics. However, with the increased availability of remotely sensed information on land use, cloud cover and rain fields (Stewart and Finch, 1993; Held, 2004) and improved macro-scale hydrological models (Vörösmarty and Moore, 1991; Vörösmarty et al., 1991; Van der Weert, 1994; Tachikawa et al., 1999), the question of 'highland–lowland interaction' may be coming closer to an answer in the not too distant future. Future modelling exercises in the region could concentrate on such comparatively data-rich basins as the Mahaweli, Sri Lanka (Madduma Bandara and Kuruppuarachchi, 1988), the Konto, East Java (Rijsdijk and Bruijnzeel, 1990, 1991), the Citarum, West Java (Van der Weert, 1994), and the Chao Phraya, Thailand (Dyhr-Nielsen, 1986; Alford, 1992; Tachikawa et al., 1999).

As for sediment delivery patterns related to land-cover transformations in the tropics, suffice to say here that there is an increasing awareness of the need to integrate on-site and off-site observations (cf. Penning de Vries et al., 1998). For too long, measurements of on-site erosion were made exclusively by agronomists whereas the determination of catchment sediment yields was usually the responsibility of river engineers. However, in recent years there have been a number of attempts (mostly by physical geographers) to bridge the gap, including several studies in Java (Bons, 1990; Rijsdijk and Bruijnzeel, 1990, 1991; Purwanto, 1999; Van Dijk, 2002) and East Malaysia (Malmer, 1990, 1996; Greer et al., 1995; Chappell et al., 1999, 2004; Douglas et al., 1999). These studies have identified trails, roads and settlements, as well as the risers of rainfed terraces (where applicable) as major sources of sediment production in many cases. Considerable progress has also been made in recent years with the formulation of physically-based on-site erosion and deposition models, some of which have been applied successfully to humid tropical conditions (Rose, 1993; Rose and Yu, 1998; Van Dijk and Bruijnzeel, 2003; Van Dijk, 2002; cf. Muzoz-Carpena et al., 1999). The next step is to link such models to distributed catchment hydrological models to enable the spatial prediction of sites with net erosion and net deposition (Vertessy et al., 1990; De Roo, 1993). Once calibrated for a given situation, such models may then be helpful in predicting the off-site effects of soil conservation measures at increasingly large distances downstream. However, in view of the rapidly increasing predictive capacity of the models, which in fact threatens to exceed our ability to check the predictions in the field, it is important to maintain a vigorous experimental field program against which model predictions can be compared in order to retain a realistic perspective (cf. Philip, 1991).

8. Conclusions

Available evidence indicates effects of forest disturbance and conversion on rainfall will be smaller in southeast Asia than the average decrease of 8% predicted for complete conversion to grassland because the radiative properties of secondary regrowth quickly resemble those of original forest. And, under 'maritime' climatic conditions, effects of land-cover change on climate will likely be less pronounced than those of changes in sea-surface temperatures.

Total annual water yield appears to increase with the percentage of forest biomass removed, but actual amounts differ between sites and years due to differences in rainfall and degree of surface disturbance. If surface disturbance remains limited, most of the water yield increase occurs as baseflow (low flows), but in the longer term rainfall infiltration is often reduced to the extent that insufficient rainy season replenishment of groundwater reserves results in strong declines in dry season flows. Although reforestation and soil conservation measures can reduce enhanced peak flows and stormflows associated with soil degradation, there is no well-documented case of a corresponding increase in low flows. While this may reflect higher water use of newly planted trees, cumulative soil erosion during the post-clearing phase may have reduced soil water storage opportunities too much for remediation to have a net positive effect in particularly bad cases.

A good plant cover can generally prevent surface erosion, and a well-developed tree cover may also reduce shallow landsliding, but more deep-seated (>3 m) slides are determined rather by geology and climate. Catchment sediment yield studies in southeast Asia demonstrate very considerable effects of such common forest disturbances as selective logging and clearing for agriculture or plantations, and, above all, urbanisation, mining and road construction.

The 'low flow problem' is the single most important 'watershed' issue requiring further research, along with evaluation of the time lag between upland soil conservation measures and any resulting changes in sediment yield at increasingly large distances downstream. Such research should be conducted within the context of the traditional paired catchment approach, complemented with process-based measuring and modelling techniques. More attention should also be paid to underlying geological controls of catchment hydrological behaviour when analysing the effect of land use change on (low) flows or sediment production.

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