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LETTER

The effects of annual precipitation and mean air temperature on annual runoff in global forest regions

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Abstract Changing trends in runoff and water balance under a warming atmosphere are a major subject of interest in recent climatic and hydrological research. Forest basins represent the most complex systems including critical hydrological processes. In this study, we investigate the relationship between annual total runoff (Q), precipitation (P), and mean temperature (T) using observed data collected from 829 (forest) site years around the world. It is shown that the strong linear relationship between annual P and Q is a function of mean T. By empirically perturbing observed annual Q and P with T, a set of ΔQ-zero lines are derived for different mean T. To evaluate the extent to which the future changes in annual P and T alter Q, the future projections of ΔP and ΔT under a warming scenario (A1B) from five coupled AOGCMs (Atmosphere-Ocean General Circulation Models) are compared with the empirical ΔQ-zero lines derived in this study. It is found that five AOGCMs show different distributions with respect to the ΔQ-zero lines, which can be attributed to the contrasting dominant sensitivities of various influencing factors to water balance partitioning among models. The knowledge gained in this empirical study is helpful to predict water resources changes under changing climate as well as to interpret hydrologic simulations in AOGCM future projections.

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1 Introduction

Nearly 4 billion hectares of forests (~30% of the total land area) cover the Earth's surface. Forest biomes represent the most hydrologically complex ecosystems since the water balance of forest ecosystems is influenced not only by the near-surface hydrologic processes, also by the physiology of diverse forest plants. Further, the factors affecting the near-surface boundary conditions (e.g., albedo, roughness length, soil temperature, etc.) of forest are more variable than other biomes (Komatsu 2005; Bonan 2008; Cho et al. 2011). Forests provide a significant contribution to the regional and global water cycle, consuming more water than other terrestrial biomes (Kanae et al. 2001; Bonan 2008). The distribution of precipitation (P) in forests through the year varies widely according to climate zones and land elevation. Evapotranspiration (E) from the forest returns a large amount of surface water to the atmosphere with its variation highly sensitive to climatic and surface conditions (Turner 1991; Bonan 2008). The pioneering study of forest hydrology conducted by Hibbert (1967) found that the hydrological response of forests to climate forcing is highly variable and difficult to predict. Therefore, investigations of the hydrology in a forest environment deserve more attention in the context of global hydrology and water resource management.

Air temperature (T) is a critical climatic variable for hydrologic and physiological processes in the terrestrial water and energy cycle (Jarvis et al. 1976; Brutsaert 1982; Jones 1992; Campbell and Norman 1998). Most observations and climate modeling assessments indicate that T has increased over the 20th century (Trenberth et al. 2007) and will continue to increase in this century (Randall et al. 2007). A warmer atmosphere is generally accompanied with higher P on the basis of the Clausius-Clapeyron relation, which states that the atmosphere water vapor pressure has a positive exponential relationship with T. Increased surface runoff (Q) is expected in accordance with increased P. Furthermore, both the increased surface available energy (i.e., the sum of latent and sensible heat fluxes) and length of growing season under rising T will enhance E, resulting in an increase in atmospheric water vapor and cloud liquid water that are the resources that subsequently fall as precipitation. On the other hand, land surface water loss due to enhanced E may cause a decline in Q, particularly for the forest where typically E is relatively high.

Wetherald and Manabe (2002), using an ensemble analysis of 20 coupled ocean-atmosphere-land model simulations up to the year 2050, reported the global increase in P, E, and Q of 5.2, 5.2, and 7.3%, respectively, as a result of CO_2-induced increase in surface T of 2.3°C. In addition, an increasing trend in future P projections over the global land and ocean has been simulated by most of the participating models in the Program for Climate Model Diagnosis and Intercomparison (PCMDI) (Meehl et al. 2005; Nohara et al. 2006; Waliser et al. 2007), which it can be anticipated by the Clausius-Clapeyron relation. A warmed atmosphere will also further increase the rate of E (Campbell and Norman 1998). However, the extent to which E is changed, and even the trend direction, in response to higher T is uncertain among model projections of the future warming scenarios (Nohara et al. 2006). Further, studies based on long-term observation records or climate modeling have recently attributed the decreasing of E due to solar dimming (Roderick and Farquhar 2002), soil moisture shortage (Jung et al. 2010), deforestation (Bradshaw et al. 2007), or CO_2-induced suppression of plant transpiration (Gedney et al. 2006). Increased P together with decreased E could cause an increase in Q.

The change in terrestrial water balance under altered environmental conditions has a critical effect on sustainable water management for human activity, particularly the case in Q (Oki and Kanae 2006). Some future climate modeling studies as well as observational data analysis show that the annual Q has increased in many regions of the world in the 20th century (Probst and Tardy 1987; Labat et al. 2004). However, the environmental factors responsible for the increase in Q is still not clear (Labat et al. 2004; Legates et al. 2005; Huntington 2006). To understand the change in annual forest Q, the changes in the basic climatic factors, particularly P and T, associated with Q should be the priority to investigate. In this study we investigate the relationships among annual Q, P, and T using observational data at various sites around the world, and examine the extent to which changes in individual P and T are required to increase or decrease Q. We expect that the information gained will be useful to understand water resources changes under climate warming scenarios as well as to compare with the hydrologic simulations of climate model projections.

2 Data and methodology

2.1 Observational data

Runoff can be derived from the general terrestrial water balance equation as follows:

$$Q = P - E - \Delta S - D \tag{1}$$

where ΔS is the change in soil water storage, and D is deep percolation to groundwater. The usefulness of water supplies for agricultural irrigation and domestic use depends on the total amount, availability, and reliability of Q. Generally, the amount of available water (Q) as regional or global water resources is evaluated by the use of annual water balance (Zhang et al. 2001; Budyko 1974) in which Q is mainly governed by the variation in P and E because the magnitude of ΔS and D can reasonably be assumed negligible in annual water balance (Zhang et al. 2001; Oki and Kanae 2006; Komatsu et al. 2008), particularly for large natural river basins without significant human influences (Brutsaert 1982; Milly 1994).

The observed annual Q and P are the common hydrologic quantities in water budget studies in forest basins. T is a conventionally measured climatological variable at most climate stations. Most previous studies on the synthesis of global Q were based on only a limited number of Q measurements (Pekarova et al. 2003). By searching from the literature, we have compiled the annual Q, P, and annual mean T over 829 (forest) site years from the field measurements across various climatic zones (see Table S1 in the electrical supplement material), most of them located in large natural basin. Further, when it is possible to take annual E in the same literature, we have also compiled it as well as annual Q, P, and T. The locations of measurement sites of our compiled data are fairly uniformly distributed over each continent (see Fig. 1). Global forested regions are indicated in Fig. 1 using by the GLC2000 global land cover data for the year 2000 (Bartholomé and Belward 2005). The distribution of annual E on Fig. 1 shows the higher annual E in the tropical areas and lower E in the high-latitude areas. Without the efforts of making any additional analyses, the values of annual Q, P, T, and E as reported in the individual literature are used in this study, although the methods of measurement and estimation may differ among different forest sites.

2.2 Regression relationship between annual P and Q

Annual P in the forest regions varies both temporally and spatially and is the largest term in the annual water balance. The runoff coefficient K (the ratio of Q to P amount) is widely

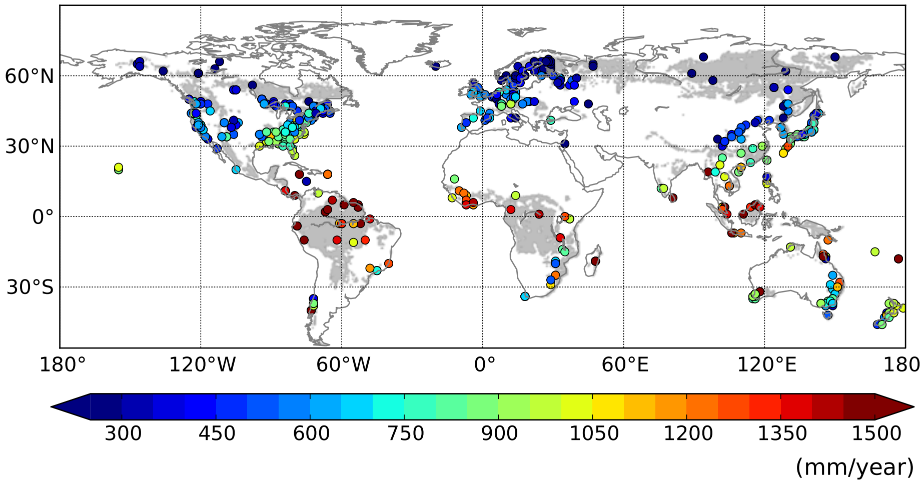


Fig. 1 Locations of observational data used in this study. Gray areas indicate forest regions according to GLC2000 land cover map at the 1-km resolution (Bartholomé and Belward 2005). The color bar indicates the values of annual total evapotranspiration (E)

used as a fundamental hydrologic index to characterize and classify catchments (Savenije 1996; Brutsaert 2005). K is a function of both surface condition factors (e.g., forest type, forest harvesting, soil type, watershed topographic slope, stream network, catchment area, land practice, etc.) and climate factors (e.g., rainfall magnitude and frequency, amount of sunshine, temperature, etc.) (Kadioglu and Şen 2001; Chen et al. 2007). Among various climate factors, T is most commonly related to K in the previous annual water balance studies with in general a negative relationship between them (e.g., Chen et al. 2007; Vlčková et al. 2009).

The relationship between annual P and Q is generally non-linear. Due to the complexity of forest systems, however, a commonly accepted mathematical models on the P-Q relationship has yet to be established, although some previous studies used the non-linear regression approach to explore their relationship (e.g., Hudson 1993; Kadioglu and Şen 2001; Partal 2010). In fact, P has a first-order control on the large-scale forest Q (e.g., Hibbert 1967; Andréassian 2004; Komatsu et al. 2011). In this study, annual Q is assumed to change linearly with annual P. The conventional linear regression approach is employed to investigate the P-Q relationship using the compiled annual P and Q data stratified by annual mean T. Surface condition factors and other meteorological factors are assumed to have minor influences on the annual P-Q relationship.

2.3 Future projections from climate modelling

To characterize future changes in P and T, we use the global simulation outputs for two years (2001 and 2100) from five different coupled AOGCMs (Atmosphere-Ocean General Circulation Models), which were archived in the datasets of the A1B scenario (i.e., a balanced emphasis across all energy sources, in more integrated world) simulations at PCMDI (see Table S2 in the electrical supplement material). While these models do not include groundwater, biological water, lake, and river storage, the variables directly associated with the global water cycle (e.g., precipitation, precipitable water, runoff, cloud water over ocean, soil moisture, and snow mass) are estimated in their simulations (Waliser et al. 2007). All simulated results are converted to a common 1° by 1° grid by using a bicubic spline interpolation scheme.

3 Results and discussion

3.1 General relation between annual P and Q as a function of T

T reasonably performs the sensitivity analysis in surface water partitioning to P (in long-term time scale) because of the relationship between a higher T strongly causes an increase in potential E (Brutsaert 1982), despite the fact that the factors in decline of E are reported recently (see section 1). Our compiled data also confirm the positive trend of E to T (see Fig. S1 in the electrical supplement material). However, when annual P is small, a lower E and decreased E-T sensitivity could be found due to water limitation. Based on these relationships among annual P, Q, E and T, we attempt to formulate empirically a general relationship between P and Q as a function of T.

Figure 2 plots the relationship between annual total P and Q as a function of annual mean T. Generally, a clear linear relationship between P and Q is observed. The linear regression between P and Q can be classified by the annual mean T: $Q = 0.82P - 162.9(R^2 = 0.77)$ for −10~0°C, $Q = 0.88P - 384.5(R^2 = 0.93)$ for 0~10°C, $Q = 0.88P - 623.4(R^2 = 0.89)$ for 10~20°C, and $Q = 0.89P - 1039.2(R^2 = 0.90)$ for 20~30°C. The slopes of all regression lines, that is, the ratio of Q to P, are relatively constant under various forest and climatic conditions with the average value of 0.87 and a standard deviation of only 0.03. The linear regression line of P and Q at lower T is located in a higher part than that at higher T. Consequently, the intercept in the vertical axis (b) has a negative relation with T (see Fig. 3), which can be fitted by $b = -0.4855T^2 - 18.968T - 253.57$. Even though other factors such as the type of forests and surface micro-topography have an influence on Q (Hibbert 1967; Bosch and Hewlett 1982; Andréassian 2004), it is noteworthy that the slope of the P-Q relation (Fig. 2) can be distinguished by only T. This is because E, as strongly governed by T, will likely have the significant effect on the variation of Q at the annual timescale (Komatsu et al. 2008).

Fig. 2 Scatter plots between annual total precipitation (P) and runoff (Q) using observational data from various forest basins around the world (Fig. 1). Color bar indicates annual mean temperature (T). Four linear regression lines are fit for the data within the following T ranges: −10~0°C, 0~10°C, 10~20°C, and 20~30°C

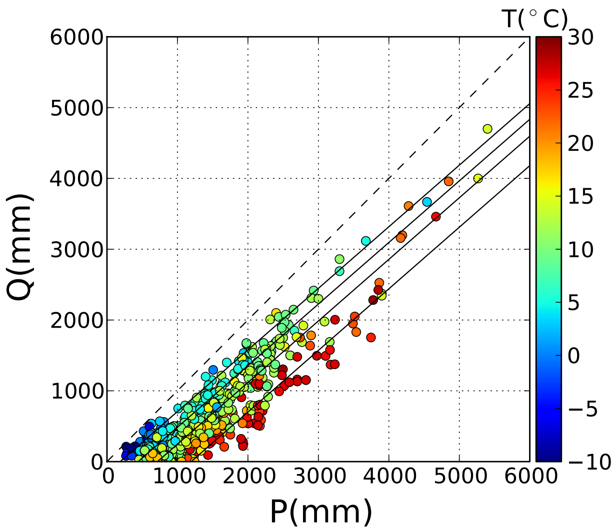
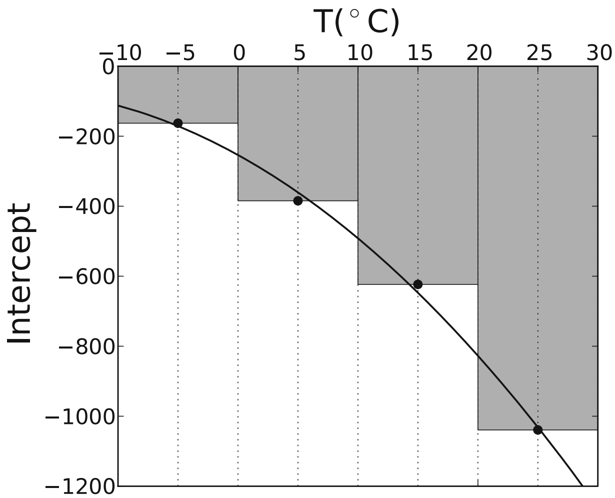


Fig. 3 Correlation between temperature and the intercepts of the linear regression lines between annual runoff and precipiation (see Fig. 2)



3.2 Change in annual Q induced by changes in annual P and T

By combining the empirical relations between P and Q derived in the section 3.1 (see Figs. 2 and 3), the following equation can be derived:

$$Q = 0.87P + \left(-0.4855T^2 - 18.968T - 253.57\right) \tag{2}$$

To investigate the extent to which the changes in annual P and T alter annual Q, given a set of perturbations of ΔP and ΔT, we assume ΔQ=0 in Eq. (2) and calculate the line between ΔP and ΔT. The derived ΔQ-zero lines are plotted in Fig. 4 for different annual mean T. From this figure, a positive relation between ΔP and ΔT can be observed with the sensitivity (slope magnitude) increasing as the initial T (i.e., T before the change) decreases. For example, when P is increased by 200 mm, T has to be increased by about 5° if ΔQ is zero. Given a set of ΔP and ΔT for initial T, the horizontal distance from the ΔQ-zero lines represents the additional precipitation (relative to the ΔQ-zero condition) for the partitioning between ΔE and ΔQ. In addition, at ΔP=200 mm again, if ΔT change larger than 5°, the annual loss of E exceeds the increase in annual P and hence ΔQ<0. That is, the farther a point is left to the ΔQ-zero lines, the larger potential for the negative ΔQ. In contrast, ΔQ will be positive if it is right to the ΔQ-zero lines. In summary, the derived ΔQ-zero lines can qualitatively be used to judge the potential influence of atmospheric warming induced increase in annual E and the associated change in annual P on the annual Q (also see Appendix A).

3.3 Applications to future P and T data

In Fig. 4, the annual ΔP and ΔT from 2001 to 2100 (i.e., the values of 2100 minus that of 2001) simulated by five AOGCMs for the global grids of forest (Fig. 1) are plotted together with the derived ΔQ-zero lines. The corresponding change in simulated annual runoff (ΔQ) is also shown in this plot. If climate change in P and T are the predominant factors for the runoff change, then the sign (and to some extent its magnitude) of ΔQ should be consistent with the ΔQ-zero lines. Accordingly, negative (positive) ΔQ, as plotted in red (blue) in Fig. 4, is expected to the left (right) of the ΔQ-zero lines. As observed from Fig. 4,

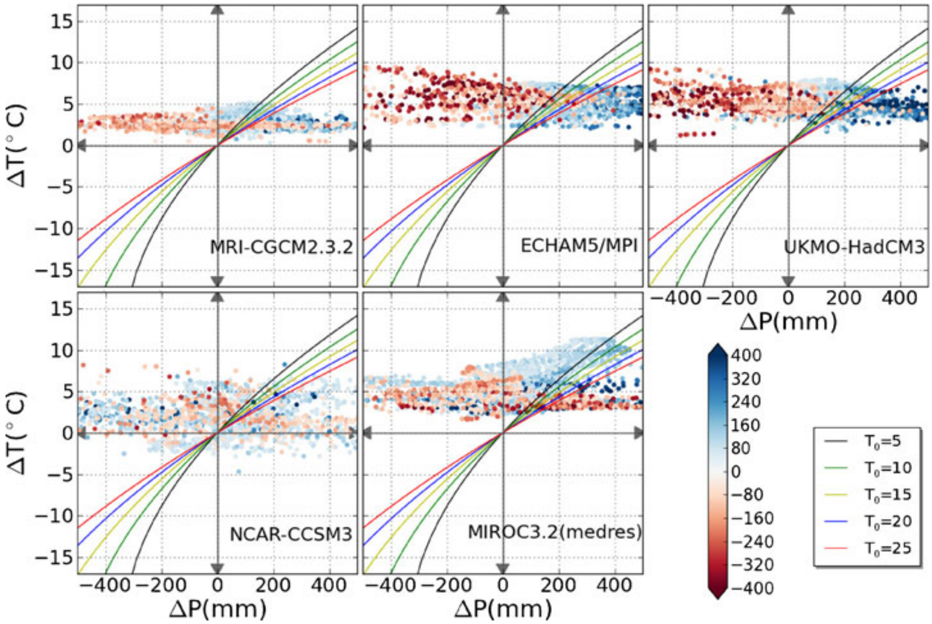


Fig. 4 The ΔQ-zero lines as a function of ΔP and ΔT (Eq. 2). See the text for explanation. The colors of lines indicate mean annual temperature T before change. Scattering plots show the data ΔP and ΔT from 2001 to 2100 simulated by five AOGCMs using the projected A1B scenario. The color bar indicates model-simulated annual runoff change ΔQ

however, five AOGCMs differ significantly in the prediction of the future hydrological change under climate warming. The ΔQ simulated by MRI-CGCM2.3.2, ECHAM5/MPI, and UKMO-HadCM3 are generally more consistent with the above expectation than the other two models, with most of their $\Delta Q(+)$ ($\Delta Q(-)$) grids located on the right (left) to the ΔQ-zero lines (see Fig. 4). It implies that hydrologic partitioning in these three AOGCMs is highly controlled by the change of P and T, while other factors are likely to only have secondary controls on simulated runoff changes. For the NCAR-CCSM3, the distribution of $\Delta Q(+)$ and $\Delta Q(-)$ is mixed together and lacks of any identifiable patterns. For the MIROC3.2(medres), although the cluster of $\Delta Q(+)$ and $\Delta Q(-)$ can be clearly differentiated in that most of the $\Delta Q(+)$ grids has a larger $\Delta T(+)$ than the $\Delta Q(-)$ grids, they are located to the left of the ΔQ-zero lines, which is in contrast with the other three models (MRI-CGCM2.3.2, ECHAM5/MPI, and UKMO-HadCM3). This discrepancy mostly occurs in the high latitude areas ΔT has the largest increase and snow is the critical factor in terrestrial water cycle (see Fig. S2 in the electrical supplement material).

The reasons for the discrepancy found in the above comparison with AOGCM simulations are mainly due to the fact that our empirical ΔQ-zero lines do not include any other factors with discernible effects on the change in Q than climate change (e.g., parameterization related to forest types and soil types, precipitation density, growth season length, the sensitivity of CO_2 reduced transpiration, etc.). For example, Gerten et al. (2008) using an offline-simulation of a global vegetation and hydrological model over the period 1901–2002 demonstrated that the change in land cover, in addition to their independent effect of increasing P, is a more significant factor in the decrease of Q than the rising temperature. Also, observational data have shown that when the deforestation areas increase, the corresponding Q increases (e.g., Stednick 1996; Bradshaw et al. 2007).

4 Summary and conclusions

On an annual basis, surface Q generally has a close correlation with annual P (Budyko 1974). However, P is not evenly distributed around the world; about 60% of annual land P is returned to the atmosphere by global land E (Oki and Kanae 2006). Moreover, the pattern of P and E will fluctuate with climate warming. Held and Soden (2006) expect that wet (dry) regions will get wetter (drier) due to the proportional relationship between E and P. For this reason, the change in trends in Q via water balance under warmed atmosphere conditions has recently become an object of interest in climatic and hydrological research. However, the change in global Q is still poorly understood, as are the various influences on the trends in annual Q. In addition, water cycle processes in forest regions, which are characterized by P and T, are difficult to analyze due to the biophysical and physiological complexity of forest systems.

In this study, observed data of annual P, Q, E and T for the totally 829 forest site years are used to derive an empirical equation, the ΔQ-zero line between ΔP and ΔT which explain the variation in Q under the present climate. Indeed, an investigation of the effects of P and T is required prior to quantifying the effect of other discernible factors in Q as P and T are expected to largely change the climate warming situation. Therefore, our empirical the equation is applicable to predictions of Q under changing climate, we examined the conditions of P and T required for increase or decrease of Q. The results of this examination might be different from those derived from AOGCM simulations and observations under changing climate, because the examinations only consider temporal changes in P and T. However, the knowledge gained in this empirical study is helpful to predict water resources changes under changing climate as well as to interpret hydrologic simulations in climate model future projections.

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Appendix A

The zero-lines of ΔQ (= $Q_1 - Q_0$) between ΔP (= $P_1 - P_0$) and ΔT (= $T_1 - T_0$) in Fig. 4 are calculated from Eq. (2).

$$\begin{aligned} Q_0 &= 0.87P_0 + \left(-0.4855T_0^2 - 18.968T_0 - 253.57\right) \\ Q_1 &= 0.87P_1 + \left(-0.4855T_1^2 - 18.968T_1 - 253.57\right) \end{aligned} \tag{A1}$$

When ΔQ is zero, the following equation can be derived:

$$\Delta T = \frac{0.0459}{(1 - 0.0256(T_1 + T_0))}\Delta P \tag{A2}$$

Thus, the ΔQ-zero line has a positive relationship between ΔP and ΔT, and it depends on T_0, but not on P_0.

References

Andréassian V (2004) Waters and forests: from historical controversy to scientific debate. J Hydrol 291:1–27

Bartholomé E, Belward AS (2005) GLC2000: a new approach to global land cover mapping from Earth observation data. Int J Remote Sens 26:1959–1977

Bonan GB (2008) Forests and climate change: forcings, feedbacks, and the climate benefits of forests. Science 320:1444–1449

Bosch JM, Hewlett JD (1982) A review of catchment experiments to determine the effect of vegetation changes on water yield and evapotranspiration. J Hydrol 55:3–23

Bradshaw CJA, Sodhi NS, Peh KS-H, Brook BW (2007) Global evidence that deforestation amplifies flood risk and severity in the developing world. Glob Change Biol 13:2379–2395. doi:10.1111/j.1365-2486.2007.01446.x

Brutsaert W (1982) Evaporation into the atmosphere; theory, history, and application. Kluwer, Dordrecht

Brutsaert W (2005) Hydrology: an introduction. Cambridge University Press, Cambridge

Budyko MI (1974) Climate and life. Academic, New York

Campbell GS, Norman JM (1998) An introduction to environmental biophysics. Springer, New York

Chen L, Liu C, Li Y, Wang G (2007) Impacts of climatic factors on runoff coefficients in source regions of the Huanghe River. Chin Geogr Sci 17:47–55

Cho J, Miyazaki S, Yeh PJ-F, Kim W, Kanae S, Oki T (2011) Testing the hypothesis on the relationship between aerodynamic roughness length and albedo using vegetation structure parameters. Int J Biometeorol doi:10.1007/s00484-011-0445-2 (In press)

Gedney N, Cox PM, Betts RA, Boucher O, Huntingford C, Stott PA (2006) Detection of a direct carbon dioxide effect in continental river runoff records. Nature 439:835–838

Gerten D, Rost S, von Bloh W, Lucht W (2008) Causes of change in 20th century global river discharge. Geophys Res Lett 35:L20405. doi:10.1029/2008GL035258

Held IM, Soden BJ (2006) Robust responses of the hydrological cycle to global warming. J Climate 19:5686–5699

Hibbert AR (1967) Forest treatment effects on water yield. In: Sopper WE, Lull HW (eds) Forest Hydrology, Proceedings of a National Science Foundation Advanced Science Seminar. Pergamon Press, Oxford

Hudson NW (1993) Field measurement of soil erosion and runoff. FAO soils bulletin 68. Food and Agriculture Organization of the United Nations, Rome

Huntington TG (2006) Evidence for intensification of the global water cycle: review and synthesis. J Hydrol 319:83–95

Jarvis PG, James GB, Landsberg JJ (1976) Coniferous forest. In: Monteith JL (ed) Vegetation and the atmosphere, vol 2. Academic, London

Jones HG (1992) Plants and microclimate: a quantitative approach to environmental plant physiology. Cambridge University Press, New York

Jung M, Reichstein M, Ciais P, Seneviratne SI, Sheffield J, Goulden ML, Bonan G, Cescatti A, Chen J, de Jeu R, Dolman AJ, Eugster W, Gerten D, Gianelle D, Gobron N, Heinke J, Kimball J, Law BE, Montagnani L, Mu Q, Mueller B, Oleson K, Papale D, Richardson AD, Roupsard O, Running S, Tomelleri E, Viovy N, Weber U, Williams C, Wood E, Zaehle S, Zhang K (2010) Recent decline in the global land evapotranspiration trend due to limited moisture supply. Nature 467:951–954

Kadioglu M, Şen Z (2001) Monthly precipitation-runoff polygons and mean runoff coefficients. Hydrol Sci J 46:3–11

Kanae S, Oki T, Musiake K (2001) Impact of deforestation on regional precipitation over the Indochina Peninsula. J Hydrometeorol 2:51–70

Komatsu H (2005) Forest categorization according to dry-canopy evaporation rates in a growing season: comparison of the Priestley-Taylor coefficient values from various observation sites. Hydrol Process 19:3873–3896

Komatsu H, Maita E, Otsuki K (2008) A model to estimate annual forest evapotranspiration in Japan from mean annual temperature. J Hydrol 348:330–340

Komatsu H, Shinohara Y, Kume T, Otsuki K (2011) Changes in peak flow with decreased forestry practices: analysis using watershed runoff data. J Environ Manag 92:1528–1536

Labat D, Goddéris Y, Probst JL, Guyot JL (2004) Evidence for global runoff increase related to climate warming. Adv Water Resour 27:631–642

Legates DR, Lins HF, McCabe GJ (2005) Comments on “Evidence for global runoff increase related to climate warming” by Labat et al. Adv Water Resour 28:1310–1315

Meehl GA, Covey C, McAvaney B, Latif M, Stouffer RJ (2005) Overview of the coupled model intercomparison project. Bull Am Meteorol Soc 86:89–93

Milly PCD (1994) Climate, soil water storage, and the average annual water balance. Water Resour Res 30:2143–2156

Nohara D, Kitoh A, Hosaka M, Oki T (2006) Impact of climate change on river discharge projected by multimodel ensemble. J Hydrometeorol 7:1076–1089

Oki T, Kanae S (2006) Global hydrological cycles and world water resources. Science 313:1068–1072

Partal T (2010) Wavelet analysis and multi-scale characteristics of the runoff and precipitation series of the Aegean region (Turkey). Int J Climatol. doi:10.1002/joc.2245

Pekarova P, Miklanek P, Pekar J (2003) Spatial and temporal runoff oscillation analysis of the main rivers of the world during the 19th–20th centuries. J Hydrol 274:62–79

Probst JL, Tardy Y (1987) Long range streamflow and world continental runoff fluctuations since the beginning of this century. J Hydrol 94:289–311

Randall D, Wood RA, Bony S, Colman R, Fichefet T, Fyfe J, Kattsov V, Pitman AJ, Shukla J, Srinivasan J, Stouffer RJ, Sumi A, Taylor K (2007) Climate models and their evaluation. In: Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, Miller HL (eds) Climate change 2007: the scientific basis. Contribution of working group I to the fourth assessment report of the intergovernmental panel on climate change. Cambridge University Press, New York

Roderick ML, Farquhar GD (2002) The cause of decreased pan evaporation over the past 50 years. Science 298:1410–1411

Savenije HHG (1996) The runoff coefficient as the key to moisture recycling. J Hydrol 176:219–225

Stednick JD (1996) Monitoring the effects of timber harvest on annual water yield. J Hydrol 176:79–95

Trenberth K, Jones PD, Ambenje P, Bojariu R, Easterling D, Klein Tank A, Parker D, Rahimzadeh F, Renwick JA, Rusticucci MBS, Zhai P (2007) Observations: surface and atmospheric climate change. In: Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, Miller HL (eds) Climate change 2007: the physical science basis. Contribution of working group I to the fourth assessment report of the intergovernmental panel on climate change. Cambridge University Press, Cambridge

Turner KM (1991) Annual evapotranspiration of native vegetation in a Mediterranean-type climate. Water Resour Bull 27:1–6

Vlčková M, Nechvátal M, Soukup M (2009) Annual runoff coefficient in the Cerhovický stream catchment. J Water Land Dev 13b:41–56

Waliser D, Seo K-W, Schubert S, Njoku E (2007) Global water cycle agreement in the climate models assessed in the IPCC AR4. Geophys Res Lett 34:L16705. doi:10.1029/2007GL030675

Wetherald RT, Manabe S (2002) Simulation of hydrologic changes associated with global warming. J Geophys Res 107(D19):4379. doi:10.1029/2001D001195

Zhang L, Dawes WR, Walker GR (2001) Response of mean annual evapotranspiration to vegetation changes at catchment scale. Water Resour Res 37:701–708