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Modifiers and Amplifiers of High and low Flows on the Ping River in Northern Thailand (1921–2009): The Roles of Climatic Events and Anthropogenic Activity

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Abstract We analyse an 89-year streamflow record (1921–2009) from the Upper Ping River in northern Thailand to determine if anomalous flows have increased over time (Trenberth, Clim Res 47:123–138, 1999; Trenberth, Clim Chang 42:327–339, 2011). We also relate the temporal behavior of high and low flows to climatic phenomena and anthropogenic activities. Peak flows have not increased significantly since 1921. However, minimum flows showed a very significant downward trend over the study period (α=0.01). Annual and wet season discharge show significant downward trends (α=0.05). All flow variables appear to be more variable now than 90 years ago especially annual peak flows. Both annual peak and minimum flows are correlated with annual and wet season rainfall totals. Minimum flow is also sensitive to the length of the monsoon season and number of rainy days in the previous monsoon season. Peak flow activity is driven predominantly by climate phenomena, such as tropical storm activity and monsoon anomalies, but the relationship between peak flows and ENSO phenomena is unclear. In general, annual discharge variables did not correspond unequivocally with El Ninõ or La Ninã events. Minimum flows show a major decline from the mid-1950s in line with major anthropogenic changes in the catchment. The plausible intensification of the hydrological cycle that may accompany global warming is of concern because of the potential to affect tropical storm activity and monsoon anomalies, phenomena that are linked with very high flows in this river system. The obvious effect of human activities such as reservoir management on low flows calls for careful management to prevent droughts in the future.

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1 Introduction

Although floods and droughts are common in Monsoon Asia, there is a growing public perception they are increasing in magnitude and frequency due to a multitude of factors, including climate change, deforestation, urbanization, and population growth (cf. Forsyth and Walker 2008; van Dijk et al. 2009). Land-cover changes in the Southeast Asian region are already believed to have changed the regional monsoon flow patterns that affect hydro-climatological variables including streamflow (Kanae et al. 2001; Sen et al. 2004). Climate change scientists note the possibility of an increase in extreme floods due to increases in rainfall frequency and intensity that may accompany acceleration in the hydrological cycle; meanwhile reductions in rainfall or increased evapotranspiration could cause low stream flows or droughts (Trenberth 1999, 2011).

Public sentiment that deforestation causes floods, however, is not supported by research on large catchments (cf. Bruijnzeel 2004; Bradshaw et al. 2007; Forsyth and Walker 2008). Floods are triggered by extreme rainfall events, with or without forests (Hewlett 1982; FAO 2005; Saghafian et al. 2008; van Dijk et al. 2009; Bathurst et al. 2011; Sriwongsitanon and Taesombat 2011). Urban development and the general increase in the number of people living in flood-prone areas compound the consequences associated with flooding, thereby adding to perceptions of increased flooding (Laurance 2007; Lebel et al. 2009; Wheater and Evans 2009). Wide-spread increases in road networks may also affect storm peaks and dry-season flows in some catchments (Cuo et al. 2006, 2008). Also, in contrast with public sentiment is the notion that afforestation or reforestation can reduce stream flows at some spatial scales through increased evapotranspiration (Bruijnzeel 2004). Intensification in agriculture and an increased reliance on irrigation in the dry season increases the incidence of extreme low flows and stream desiccation (Ziegler et al. 2009a).

Understanding the long-term streamflow modifying factors, either natural or anthropogenic, is vital to efficient catchment water management and reducing vulnerability to both floods and droughts (Palmer et al. 2008). The 2011 Chao Phraya river flood that caused $45 billion in damage in Thailand highlights the difficulty of water management in the tropics where seasonal rainfall can be unpredictable (Ziegler et al. 2012a,b). Most published work on streamflow analysis in Thailand focuses on the impact of land-use change on catchment water balances. For example, Alford (1992) and Wilk et al. (2001), using data sets of 50 and 38 years, respectively, did not find increasing streamflow trends related to deforestation in catchments in northern and northeast Thailand (also see Cuo et al. 2006, 2008). In contrast, Petchprayoon et al. (2010) reported an increasing trend in daily runoff for a central-northern catchment using a 15-year time series. This particular data set is arguably too short for understanding fully all amplifiers of stream flow, as long time series are typically required to detect changes in flow related to climate/landuse change at spatial scales that are most important for management (cf. Ziegler et al. 2003, 2005, Assani et al. 2011).

In this analysis we examine relationships between several river discharge trends, climatic phenomena, and anthropogenic factors associated with a 89-year dataset for the upper Ping River catchment in northern Thailand. The Ping River is a major tributary to Thailand's largest river, the Chao Phraya River and is a major wet-season source of water stored in the downstream Bhumipol Reservoir, which serves irrigation and commercial users during a lengthy dry season when drought may be problematic. An understanding of the climatic and anthropogenic factors affecting high and low flows of this river system has both practical and scientific value for future water resources management in Thailand.

2 Study Site

Originating within the mountainous areas of Northern Thailand with steep hills rising to an elevation of 2,000 m with valleys below 500 m, the Ping River drains an area of 6,355 km^2 at the P1 gauging location in Chiang Mai (18◦47′09″, 99°00′29″) (Fig. 1). The width of the river at P1 is approximately 110 m; and the floodplain extends about 15 km on either side of the river bank (Wood and Ziegler 2008). The river basin is underlain by older paleozoic gneissic granites, Paleozoic sediments and volcanics, Mesozoic granitic rocks and Tertiary continental basin-fill sediments (Wood and Ziegler 2008).

The catchment was previously covered by subtropical forests that have been slowly converted to agricultural lands in the recent past due to a host of economic, social and political drivers that have commonly been driving land-cover/land-use changes through montane mainland SE Asia (cf. Fox et al. 2012). The Indochina Peninsular saw the most dramatic forest loss from 1973 to 1985, with Thailand experiencing the highest rates of deforestation (Sen et al. 2004). Ongsomwang and Rattanasuwan (2009) found that forest cover in Thailand reduced from 53 % to 32 % from 1961 to 2005. Deforestation rates only stabilised after the logging ban in 1989. The current landscape in northern Thailand consists of a mosaic of fragmented forest covers interspersed with agricultural lands and urban/

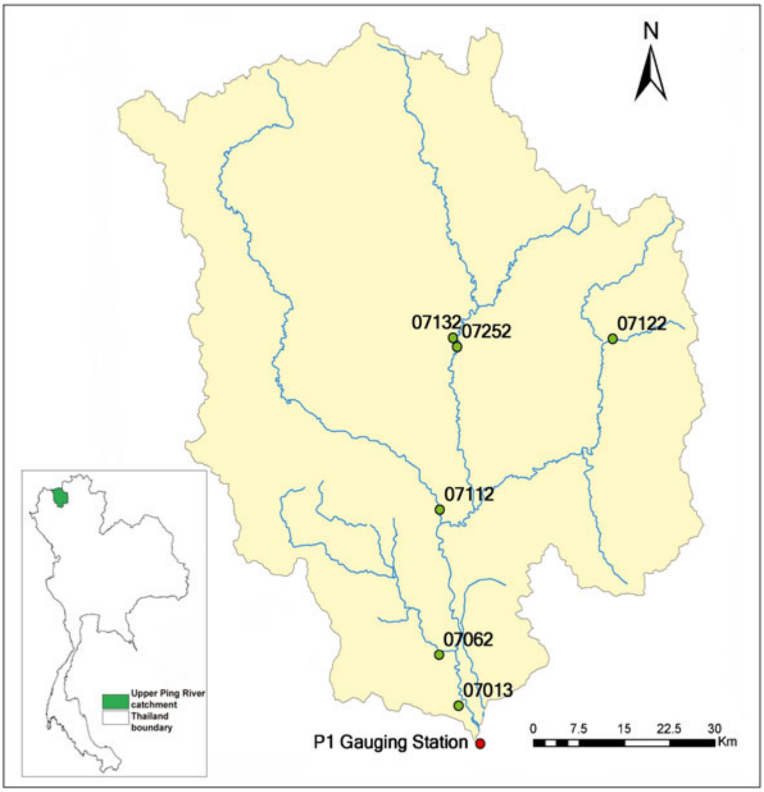

Fig. 1 Location of the Upper Ping River catchment in Thailand (insert) and rainfall stations within the catchment that have rainfall data since 1921. The *numbers* refer to the station codes

peri-urban areas (Fox et al. 2012; Ziegler et al. 2009b; Schmidt-Vogt et al. 2009; Sangawongse et al 2011).

Annual long-term rainfall in northern Thailand is 1,200 mm, with the seasonal rainfall accounting for as much as 92 % of the annual total (Wood and Ziegler 2008). Thailand has a maritime continental climate for which the large-scale climatic variability associated with the Indian Ocean and Pacific Ocean impact annual and seasonal rainfall patterns (Rasmussen and Carpenter 1983; Chen and Chappell 2009). The Southeast monsoon, extending from approximately late April or early May to October/November (Matsumoto 1997; Cook and Buckley 2009), occurs as warm air from the Indian Ocean moves northward with the Inter-Tropical Convergence Zone (ITCZ) towards Thailand. By July the ITCZ has moved to Southern China; thereby decreasing monthly rainfall. In August, the ITCZ moves southwards again over North Thailand; and peak rainfall occurs during this month or in September. The end of the monsoon period in October/November coincides with the moving of the ITCZ farther south. A dry winter season then extends until April/May of the following year (Thai Meteorological Department; http://www.tmd.go.th/en/).

Typhoons and other types of climatic disturbances in the form of monsoon depressions influence rainfall activity in the catchment. Typhoons originating from the Western North Pacific Ocean or the South China Sea pass over northern Thailand, with peak activity occurring usually from September to October (Lim and Boochabun 2012). ENSO (El Ninõ or La Niñã) events also influence the amount of rainfall falling during the wet season (Kripalani and Kulkarni 2001; Singhrattna et al. 2005a,b). El Ninõ and La Nino events are believed to correspond with dry and wet periods respectively; however, research has shown that this is not always the case (Kripalani and Kulkarni 2001; Singhrattna et al. 2005a,b).

3 Methods

3.1 Data

Herein we use daily streamflow data collected between 1921 and 2009 at the P1 gauging station to examine the following variables:

- Annual peak discharge (Q_{peak});
- Annual minimum discharge (Q_{min});
- Total annual discharge (Q_{total});
- Wet-season and dry-season discharge (Q_{wet}, Q_{dry}).

We also consider the following climate variables that modify or amplify river flows:

- Total annual, wet-season, and dry season rainfall (RF_{total}, RF_{wet}, RF_{dry});
- One- and seven-day maximum rainfall depths ($RF_{Max1\text{-}day}$, $RF_{Max7\text{-}day}$);
- Number of rain days in a year (Rain days);
- Duration of monsoon season (Monsoon length);
- Tropical storm frequency (TSfreq);
- Occurrence of ENSO events (El Ninõ or La Niñã).

Finally the following anthropogenic activities related to stream discharge are considered:

- Landcover/landuse change between 1973 and 2005;
- Development activity including urbanization, road building, and dam building.

Streamflow and climate data were obtained from the Royal Irrigation Department and the Thai Meteorological Department. The data span from 1921 to 2009 with the exception of length of monsoon season and tropical storm frequency, which are only available from 1951 to 2009. Discharge at the P1 gauging station (Fig. 1) is collected using a water level recorder which provides continuous stage readings. The stage values are converted to daily discharge values using a rating curve that is updated every year. We analyse the data with reference to the Thai water year, which starts from April and ends in March the following year. The wet season is defined as extending from April to November. This period covers the annual variability in the onset and withdrawal of the monsoon (see Matsumoto 1997; Cook and Buckley 2009). For the purpose of this study, anomalous flow or climatic conditions occur when the value exceeds ±1 standard deviation (SD or σ) from the long-term mean.

Although there are climate data for 10 rainfall stations, we use only that from Station 07013 (located in Chiang Mai city; Fig. 1), as it had the fewest data gaps (4 years). Four years with missing data were filled with data from the nearest station with valid data. Scatter in the relationships between rainfall stations prevented filling missing data from other stations within the catchment. The monsoon start and stop periods, obtained from the Thai Meteorological Department, were used to calculate the monsoon length. Tropical storms refer to typhoons (from the Western Pacific), cyclones (from the Indian Ocean) and tropical depressions. This classification is based on the definition used by the Thai Meteorological Department.

ENSO events are identified using the method developed by the Japan Meteorological Agency (JMA). The JMA Index is a 5-month running mean of spatially averaged sea-surface temperature anomalies over the tropical Pacific in the region extending 4°S–4°N and 150˚W–90˚W (ftp://www.coaps.fsu.edu/pub/JMA_SST_Index/). This method was chosen as it identifies both El Ninõ and La Ninã occurrences well (Hanley et al. 2003). An El Ninõ or La Ninã event occurs when the JMA Index is greater than 5 °C above or below the average for a least six consecutive months.

3.2 Trend Analysis

To test for trend significance in streamflow and climate time series, we use the Mann-Kendall test (Mann 1945, Kendall 1975), a non-parametric method often used to investigate trends in hydro-climatological signals (Ziegler et al. 2005; Zheng et al. 2007; Hao et al. 2008; Saghafian et al. 2008; Kliment and Matoušková 2009; Feng et al. 2011). The technique tests whether values of a series tend to increase or decrease as time increases monotonically (Helsel and Hirsch 2002). For two-sided testing of a series with n values, the null and alternative hypotheses are expressed as:

$$\left.\begin{array}{l} H_0 : prob\left[y_j > y_i\right] = 0.5 \\ H_1 : prob\left[y_j > y_i\right] \neq 0.5 \end{array}\right\} for\ all\ t_j > t_i \tag{1}$$

where time series values are denoted by y, t is time in years: i ranges from 1,2, …, (n-1); and j ranges from 2,3,…, n. Significance can be determined from a test statistic, S, that measures the monotonic dependence of y on time:

$$S = \sum_{j=1}^{n-1} \sum_{i=j+1}^{n} SGN\left(y_j - y_i\right) \tag{2}$$

where SGN is a function that returns a value of -1, 0 or 1, which reflects the sign of the expression $y_j - y_i$.

The range of S is therefore $\pm\, n(n-1)/2$. To test significance, a statistic (Z_s) that is closely approximated by the standard normal distribution is computed as:

$$Z_s = \begin{cases} \dfrac{S-1}{\sigma_S} : S > 0 \\ 0 : S = 0 \\ \dfrac{S+1}{\sigma_S} : S < 0 \end{cases} \tag{3}$$

For which σ_S is calculated as:

$$\sigma_S = \left[\frac{n(n-1)(2n+5) - \sum_{i=1}^{n} t_i i(i-1)(2i+5)}{18}\right]^{1/2} \tag{4}$$

where t_j is the number of ties of extent i. The null hypothesis is rejected at significance level α if $|Z_s| > Z_{critical}$, where $Z_{critical}$ is the value of the standard normal distribution with an exceedance probability of $\alpha/2$. Additional details regarding the Mann-Kendall test are presented elsewhere (e.g. Helsel and Hirsch 2002).

3.3 Land-Cover Analysis

Landuse classification was conducted with Landsat images using the decision tree classification method within ENVI's interactive decision tool. All pixels were divided into four landuse categories; water bodies, forest land, cropland and barren land using information such as elevation, slope, NDVI and moisture index within ENVI's transform. Due to difficulties in distinguishing cropland and barren land during the dry season, ground truthing from high resolution satellite images was performed in Google Earth to confirm and modify the decision tree classification algorithm. This resulted in landuse maps for the periods 1973/1974, 1988/89, 2000 and 2005.

4 Results

4.1 Stream Flow Trends

The peak (Q_{peak}) and minimum (Q_{min}) discharge time series show very different temporal patterns since 1921 (Fig. 2). The 15-year running mean of Q_{peak} hovers around the long-term mean until about 1965, after which a noticeable period of high peaks occurred in the 1970s. After 1984, the 15-year running mean of Q_{peak} falls below the long-term mean (Fig. 2a). However, this period is marked by wide fluctuations in high and low peak values. Owing to this high variability, no trend has occurred since 1921 ($\tau=-0.019$, p-value=0.796; Table 1). Importantly, no statistical trend exists for the period covering the last 50–60 years when anthropogenic activity—especially deforestation—was great (Table 1, Fig. 6).

Prior to 1955, Q_{min} values were almost exclusively higher than the long-term mean (Fig. 2b). In the last 55 years, however, all but 4 years experience Q_{min} values lower than the long-term mean. The years between 1975 and 1981 with Q_{min} approximately zero represent times when streamflow was too low to be recorded. The significant decrease in minimum flow since 1921 ($\tau=-0.423$; p-value<0.0001; Table 1) is likely a result in the abrupt change in low flows occurring about 1955.

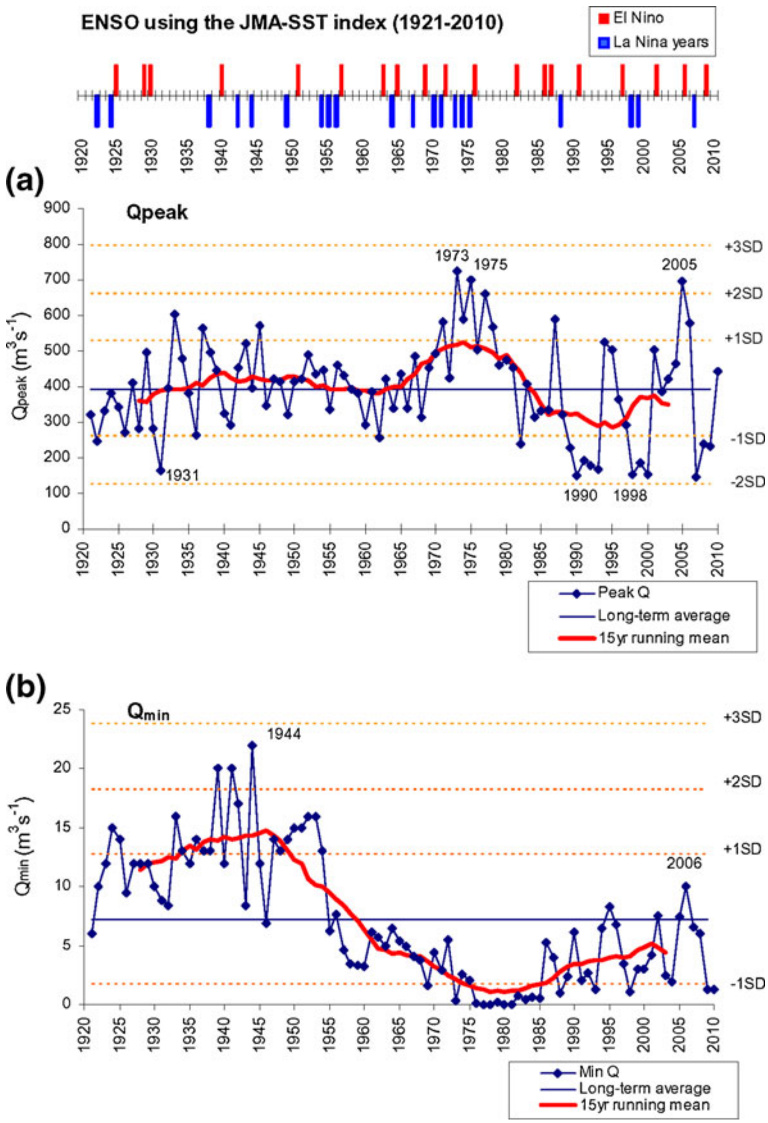

Fig. 2 Timeseries plot of (**a**) annual peak discharge, Q_{peak} and the El-Niño Southern Oscillation Index using the JMA-SST method, (**b**) annual minimum discharge, Q_{min} for the Upper Ping River catchment, 1921–2009

Annual (Q_{total}) and seasonal stream flow (Q_{wet} and Q_{dry}) time series demonstrate similar behavior, with many years having high and low extrema aligning either with Q_{peak} or Q_{min} values (Figs. 2 and 3). Despite substantial interannual variability, significant negative trends exist in the Q_{total} ($\tau=-0.202$; p-value=0.005), Q_{wet} ($\tau=-0.167$; p-value=0.021) and Q_{dry} ($\tau=-0.350$; p-value<0.0001) time series (Table 1). Furthermore, significant negative trends were detected for most dry season months (e.g., December through March), as well as wet-season months July and October (Table 1).

A summary of years where Q_{peak}, Q_{min}, and Q_{total} values exceeded 1σ above or below the long-term mean indicates that there are relatively more anomalous low flow years than high

Table 1 Descriptive statistics and trend analysis for several hydro-climatological variables for the upper Ping River (1921–2009)

Variable[a]	Units	Mean	Standard deviation	Coefficient of variation	Minimum	Maximum	Kendall's tau[b]	p value
Q_{peak}	m^3s^{-1}	394.36	134.03	0.34	146.3	726.0	−0.0189	0.796
Q_{peak} (1950–2009)	m^3s^{-1}	397.7	146.0	0.37	146.3	726.0	−0.153	0.0851
Q_{min}	m^3s^{-1}	7.21	5.51	0.77	0.0	22.00	−0.423***	< 0.0001
Q_{min} (1950–2009)	m^3s^{-1}	4.52	4.03	0.89	0.0	16.00	−0.159	0.073
Q_{total}	million m^3	1800.3	702.51	0.39	487.95	4253.9	−0.202**	0.005
Q_{wet}	million m^3	1533.3	617.16	0.40	434.94	3882.98	−0.167*	0.021
Q_{dry}	million m^3	265.5	122.32	0.46	53.0	573.0	−0.350***	< 0.0001
Apr	million m^3	38.7	21.14	0.55	1.78	125.0	−0.096	0.185
May	million m^3	79.3	40.30	0.51	10.2	208.0	0.070	0.335
Jun	million m^3	112.1	62.12	0.55	12.3	331.0	0.0051	0.946
Jul	million m^3	144.0	104.73	0.73	19.1	642.0	−0.169*	0.020
Aug	million m^3	302.7	191.96	0.63	47.9	1158.0	−0.127	0.079
Sep	million m^3	413.1	199.43	0.48	122.5	1173.0	−0.128	0.076
Oct	million m^3	278.3	120.32	0.43	26.1	610.0	−0.206**	0.0043
Nov	million m^3	167.7	81.13	0.48	32.99	401.0	−0.123	0.088
Dec	million m^3	115.8	53.64	0.46	16.7	271.0	−0.233**	0.0013
Jan	million m^3	68.9	37.79	0.55	7.15	168.0	−0.371***	< 0.0001
Feb	million m^3	42.9	25.13	0.59	5.68	123.0	−0.461***	< 0.0001
Mar	million m^3	37.8	24.59	0.65	2.41	149.0	−0.292***	< 0.0001
RF_{annual}	mm	1172.8	200.4	0.171	586.7	1780.4	−0.0904	0.211
RF_{wet}	mm	1128.2	200.0	0.177	562.2	1779.8	−0.0884	0.222
RF_{dry}	mm	43.8	41.7	0.952	0	187.4	0.0278	0.702
Max 1-day RF	mm	78.4	23.2	0.296	35.7	166.5	0.149*	0.0388
Max 7-day RF	mm	163.3	44.1	0.270	86.3	314.5	−0.0232	0.750
Rainy days	Days	108.1	19.1	0.176	36	145	0.253***	0.0004
Monsoon length	Days	137.9	37.37	0.27	25	206	0.0462	0.610
Tropical storm frequency	Integer	3.1	2.04	0.66	0	9	−0.218*	0.013

[a] Data extends only from 1951 to 2009 (n=59); n=89 for all other variables

[b] Trends are significant at α=0.05(*), 0.01 (**) or 0.001 (***)

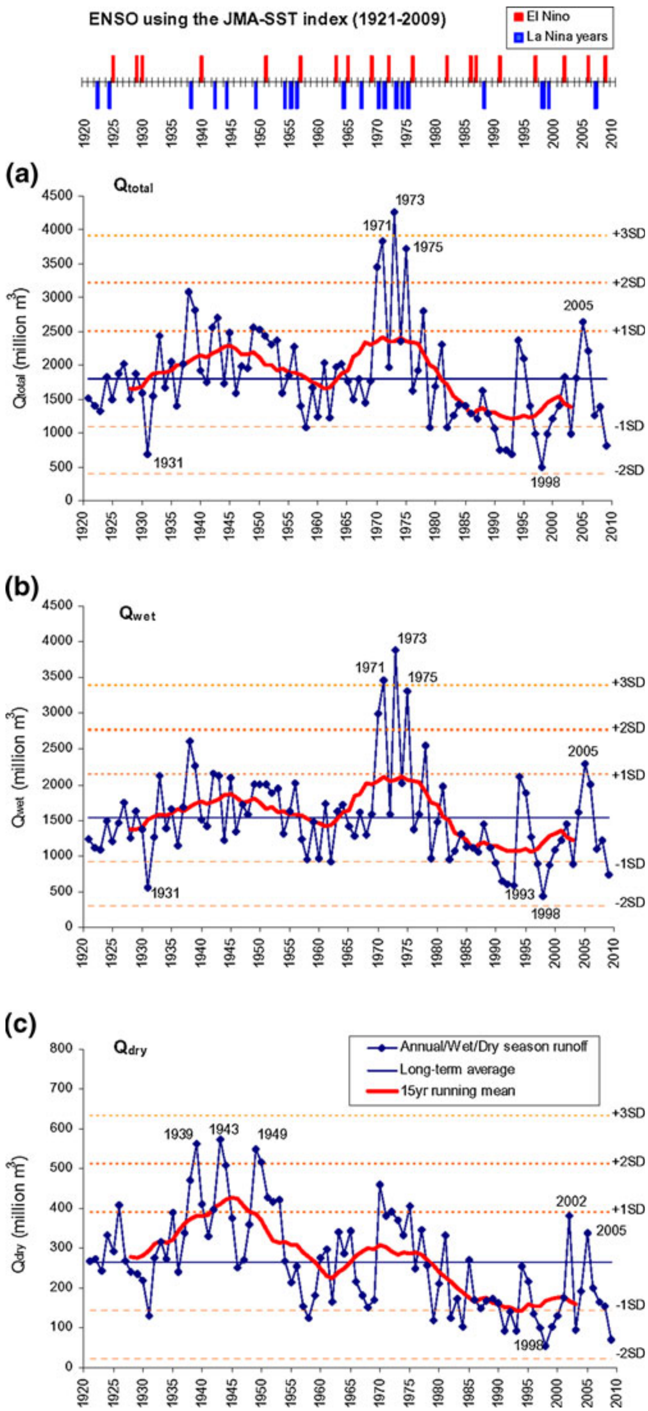

Fig. 3 Timeseries plots of runoff yields for (**a**) total annual discharge and the ENSO variability using the JMA-SST method, (**b**) wet season discharge (April–November) and (**c**) dry season discharge (December–March) for Upper Ping River catchment, 1921–2009

flow years, particularly for Q_{min} and Q_{total} (Table 2). Noticeable are the three flood years in 1973, 1975, and 2005. The 1973 peak was the highest in the 90-year record: Q_{peak} was more than 2σ above the long-term mean; and Q_{total} exceeded 3σ (Table 2). Prior to 1955, three years had Q_{min} values > 2σ above the long-term mean (Fig. 2b). Subsequently, there are 14 years since 1969 when Q_{min} was < -1σ below the long-term mean. Unlike the high flow values, the low values were never close to -2σ, but this is an artifact of the data being bound at zero.

Plots of the 15-year running mean and standard deviation show periods of changing variability in all discharge variables (Fig. 4). As stated above, Q_{peak} variability has been relatively high since 1984, but high variability exists both early in the record (1920s and 1930s) as well as in the 1970s and 1980s. Similarly, great variability occurred in the 15-year moving means of Q_{total} and Q_{wet} in the 1970s (Fig. 4c, d). By contrast, Q_{min} and Q_{dry} seem to show reduced variability over time (Fig. 4b, e). For all variables, the coefficient of variation (standard deviation/mean) for the 15-year moving mean has been increasing since 1921 (Fig. 4f). Mann-Kendall trend analysis shows statistically significant positive trends for all discharge variables at α=0.01. The observed increase in the coefficient of variation indicates greater variability in all flow variables when compared to 90 years ago.

Correlation existing among many of the various discharge variables is not unusual given dependence related to rainfall seasonality and lag effects when flows are affected by rainfall occurring in the preceding wet season (Table 3). For example, dry-season flows (Q_{dry}) are influenced by rainfall and discharge in the prior wet-season (Table 3). Similarly, wet season flows (Q_{wet}) in one year may be influenced by discharge in the previous year. Total annual rainfall and wet season rainfall is the single largest factor affecting discharge variables (Table 3). Runoff variables (Q_{peak}, Q_{total}, Q_{wet} and Q_{dry}) are significantly correlated (σ= 0.05) with annual rainfall, wet-season rainfall and maximum 7-day rainfall (Table 3). The low flow variable Q_{min} is significantly correlated with annual rainfall, wet-season rainfall, total number of rain days in a year and the length of the monsoon season (from previous monsoon season). These seasonality relationships point toward the dominance of natural over anthropogenic climatic modifiers of anomalous flows.

4.2 Climate Trends

Total (RF_{tot}) and seasonal (RF_{wet}, RF_{dry}) rainfall data showed great interannual variability, ranging from 580 to 1,790 mm, 560–1,780 mm, and 0–190 mm, for annual, wet-season, and dry-season periods (Fig. 5a, b, c). However, there were no significant trends in the annual rainfall record (Table 1). Of note was that the lowest RF_{total} in 1931 (< -3σ) corresponded with one of the lowest Q_{peak} values ever recorded (Figs. 2a, 3a and 5a). However, the greatest RF_{total} value in 1953 (> 3σ) was not associated with any high discharge value. Furthermore, the very high 7–day maximum rainfall ($RF_{7day\text{-}max}$ >3 σ) in 1967 was not associated with a high discharge peak that same year (Figs. 2a and 5f). These patterns point to spatial rainfall variability in this large catchment and the need for reliable spatially-distributed rainfall data that will allow a better understanding of climatic effects on hydrological behavior.

Trend analysis showed that one-day maximum rainfall (τ=0.149; p-value<0.0388) and number of annual rain days (τ=0.253; p-value=0.0004) have increased significantly over the last 90 years (Table 1). Although the 7–day maximum rainfall demonstrated no discernable trend, the four highest values occurred between 1965 and 1988 (Fig. 5f). These high values were largely offset by several of the lowest values, occurring in 1965, 1966, and 1984, thereby demonstrating the great variability during this 30-year period.

Table 2 Climate phenomena potentially associated with yearly high and low discharge variables

Climatic Phenomena				Qpeak			Qmin			Qtotal			
Tropical storm	Monsoon anomaly	El Nino event	La Nina event	Low ($Q_{peak} < -1\sigma$)	High ($1\sigma < Q_{peak} < 2\sigma$)	Very High ($2\sigma < Q_{peak}$)	Low ($Q_{min} < -1\sigma$)	High ($1\sigma < Q_{min} < 2\sigma$)	Very High ($2\sigma < Q_{min}$)	Low ($Q_{total} < -1\sigma$)	High ($1\sigma < Q_{total} < 2\sigma$)	Very High ($2\sigma < Q_{total < 3s}$)	Extreme ($3\sigma < Q_{total}$)
X				89, 92, 93, 00	77, 78		77, 78, 79, 80			79, 92, 93, 03	78		
	X												
		X					76						
			X	98			98			98			
X	X			90, 08		05	81, 83,85	52, 53		58, 90	05		
X		X		82, 09	87		82, 09			82, 97			
X			X		74	73	73						73
	X	X											
	X		X	99						99			
		X	X										
X	X	X			06		69	51					
X	X		X	07	71	75	88	54				70, 75	
X		X	X										
	X	X	X										
X	X	X	X										

Note: Values are years when the event occurred

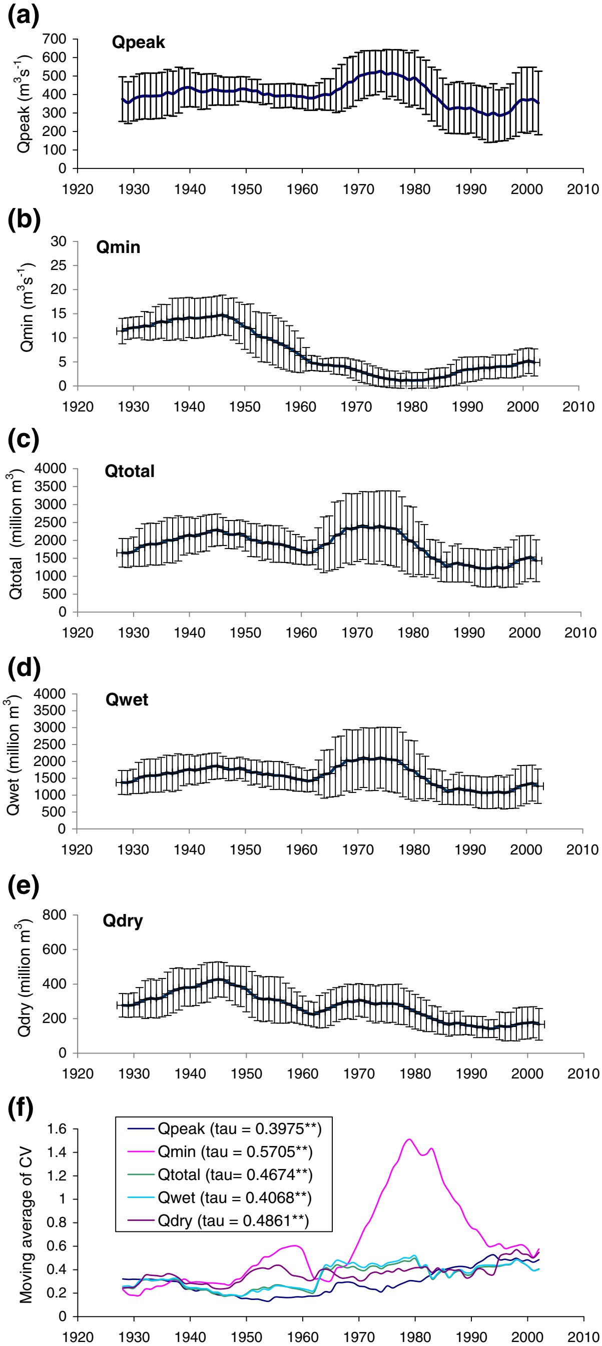

Fig. 4 Plots of the 15-year running mean discharge and standard deviation (± lSD) for (**a**) annual peak discharge, (**b**) annual minimum discharge, (**c**) total annual discharge, (**d**) wet season discharge, (**e**) dry season discharge and (**f**) 15-year running mean of the coefficient of variation (CV) for all runoff variables with Mann-Kendall tau values included (α=0.01**), Upper Ping River catchment 1921–2009

While there is no significant trend in length of monsoon season across the abbreviated time series (1951–2009), there has been a noticeable increase since 2002 (Fig. 5g). This increase, however, is similar to that which occurred from 1977 (the lowest value on record)

Table 3 Correlation (Spearman's rho) between various flow (Q), rainfall (RF), and other climatic variables

	Flow variables					Climate variables							
	Q_{peak}	Q_{min}	Q_{total}	Q_{wet}	Q_{dry}[d]	RF_{total}	RF_{wet}	RF_{dry}[d]	Raindays	$RF_{max1\text{-}day}$	$RF_{max7\text{-}day}$	Monsoon length[†]	TS_{freq}[c]
Q_{peak}	1	0.041	0.704([b])	0.725([b])	0.298([b])	0.450([b])	0.448([b])	0.147	0.06	0.066	0.345([b])	−0.158	0.026
Q_{min}[e]	0.065	1	0.416([b])	0.386([b])	0.518([b])	0.268([a])	0.299([b])	−0.015	−0.384([b])	−0.061	0.069	0.293([a])	−0.045
Q_{total}	0.704([b])	0.359([b])	1	0.987([b])	0.507([b])	0.609([b])	0.629([b])	0.074	0.097	0.065	0.337([b])	0.083	0.176
Q_{wet}	0.725([b])	0.307([b])	0.987([b])	1	0.443([b])	0.602([b])	0.629([b])	0.067	0.136	0.090	0.342([b])	0.115	0.162
Q_{dry}[f]	0.474([b])	0.535([b])	0.827([b])	0.742([b])	1	0.516([b])	0.505([b])	0.054	−0.062	−0.030	0.237([a])	−0.047	0.215

[a] Correlation is significant at $\alpha=0.05$ ($n=89$)

[b] Correlation is significant at $\alpha=0.01$ ($n=89$)

[c] Correlations with monsoon length and tropical storm frequency (TS_{freq}) are based on data from 1951–2009 ($n=59$)

[d] Correlations with RF_{dry} and Q_{dry} are based on values recorded for the previous water year because the discharge observed in the wet season of each water year (April through March the following year) is affected by the rainfall and discharge observed in the dry season of the previous year. The correlations with the other flow and climate variables are based on current water year

[e] Low flows are affected by river discharge and rainfall variables occurring during months prior to that when Q_{min} value was recorded. To carry out correlations between Q_{min} and the various variables, we had to identify when Q_{min} occurred, either in early-April at the beginning of the water year or in February/March towards the end of the same water year (i.e. next calendar year). If Q_{min} occurred in early April of a water year, the associated Q_{peak} will be from the previous water year. If Q_{min} occurs late in the water year (i.e. March), then the Q_{peak} it is correlated with is from the same water year. These adjustments were made in the raw data before correlation analysis was carried out

[f] Correlations between Q_{dry} and all flow and climate variables are based on current water year values i.e. Q_{dry} of 1999 is correlated with Q_{total} of 1999

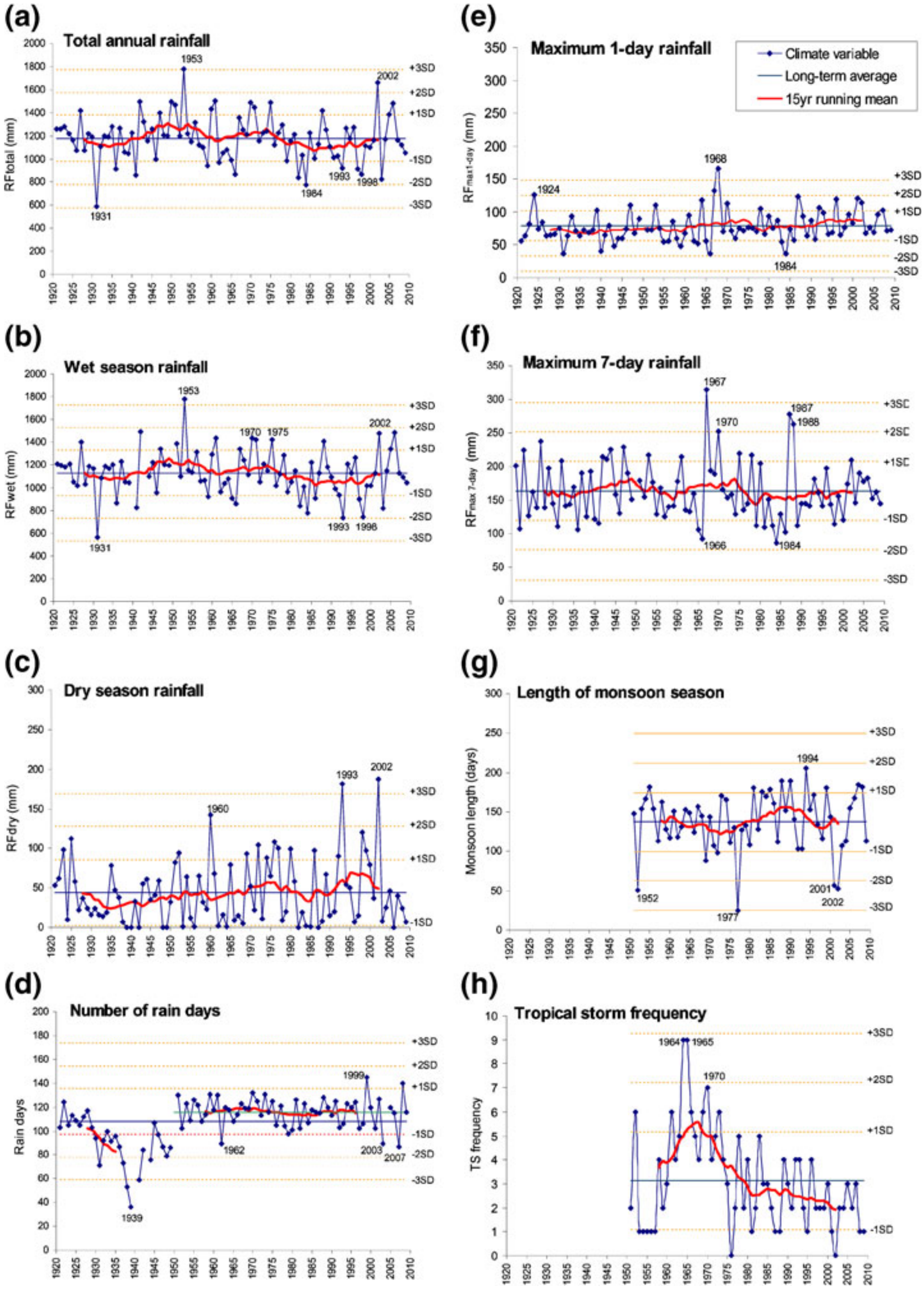

Fig. 5 Timeseries plots for climatic variables measured at Rainfall station 07013 located in Chiang Mai City (**a**–**f**, data from 1921 to 2009) and (**g**) length of monsoon season for the Northern Region (1951–2009), **f**) tropical storm frequency for Thailand (1951–2009), Source of data: Royal Irrigation Department and Thai Meteorological Department

to the mid 1980s and may reflect natural variations in long-term monsoon rainfall patterns. The variation in the length of the monsoon season experienced in northern Thailand is quite high, lasting a mere 25 days (i.e. 1977 monsoon, -3σ below the long-term mean) to over 200 days (1994 monsoon season, approximately 2σ above the long-term mean).

Tropical storm frequency has been decreasing since a high period between about 1963 and 1970 (cf. Chan and Xu 2009; Naidu et al 2011) (Fig. 5h). This trend ($\tau=-0.218$) is significant at $\alpha=0.05$. As with monsoon length, this time series extends only from 1951 to 2009. It is possible that data during the 5 year period, 1953–1957, is not reliable although data about tropical storm frequency was obtained directly from the Thai Meteorological Department. Finally, we identified 19 El Ninõ and 20 La Ninã occurrences between 1921 and 2009. The frequency of La Ninã events is higher during the period 1950–1980, while El Ninõ events appear to be increasing in frequency from 1955 (cf., Trenberth and Hoar 1997).

5 Discussion

5.1 Climatic Amplifiers and Modifiers

Kripalani and Kulkarni (1997) reported that rainfall in Asia exhibited distinct epochs of above and below normal rainfall that are not forced by ENSO events. In Thailand the epochs are supposedly on the order of 30 years. Our runoff and rainfall time series suggests the lengths of these epochs are more irregular, on the order of 15–30 years; and they are more noticeable in the Q_{total} and Q_{wet} time series than the Q_{peak} series (Figs. 2, 3 and 5). Herein, we examine 15-year running means because the period is equivalent to the low end of the observed epochs.

Our analyses also suggest the epochs somewhat correspond with ENSO events. Noticeable in the Q_{total} and Q_{wet} time series is that epochs of high and flow flows tend to be associated with the occurrence of La Ninã and El Ninõ events respectively (Fig. 3a, b). For example, the two high flow epochs (1935–1957, 1966–1980) had 6 and 7 La Ninã events compared with an associated 2 and 4 El Ninõ events (ratios of 6:2 and 7:4). Conversely, the ratio of El Ninõ to La Ninã events in the three dry epochs (before 1935, 1958–1965, 1991–present) were 2:3, 1:3, 4:8 respectively (Fig. 6).

Of the climatic phenomena potentially contributing to anomalous stream flows, the occurrence of tropical storm events stands as a major cause of runoff deviating more than 1σ above the long-term mean (Table 2). Six out of 9 years when Q_{peak} exceeded +1σ occurred in years with tropical storms occurring either with monsoon anomalies or an ENSO event (Table 2). The top three Q_{peak} values occurred in years when a tropical storm and/or a monsoon anomaly occurred. Of note is that these three high peaks occurred when more than one tropical storm struck northern and northeastern Thailand—including 2005 when three arrived late in the rainy season, causing severe flooding on the Ping River in Chiang Mai (Wood and Ziegler 2008).

The timing of tropical storm arrival within the monsoon season is also a crucial variable affecting flows. For instance, there are years when tropical storms did not produce the yearly Q_{peak} value because they arrived early in the monsoon season, probably before the basin had wetted up (e.g. during 1952 and 1982). Nevertheless, the contribution of tropical storm rainfall to annual wet season rainfall totals ranged from about 6 to 12 %—with much of the rainfall often occurring within a few days. Consequently, most of the highest annual Q_{peak} values in our 90-year record are caused by intense rainfall associated with late-season tropical storms, occurring when the catchment is wet.

The early 1970s experienced a clustering of extreme Q_{peak}, Q_{total}, and Q_{wet} values that correspond with a higher frequency of La Ninã events, tropical storm activity and monsoon

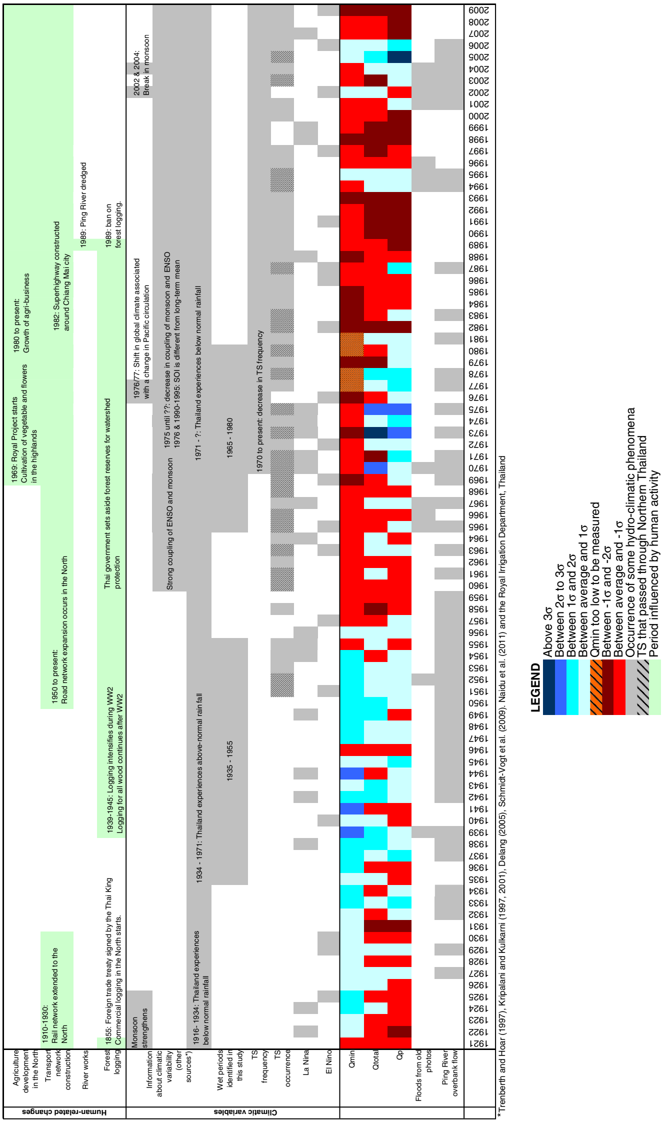

Fig. 6 Timeseries plot Q_{peak}, Q_{total} and Q_{min} together with information about climate and anthropogenic landcover changes from published sources

irregularities. The extreme weather activity in this period may have been wide-spread, as continental discharge was very high in 1972 (Dai et al 2009). By contrast, Q_{peak} was extremely low in several years in the 1990s, perhaps because of strong El Ninõ events and the 1991 eruption of Mount Pinatubo, which may have reduced rainfall (Trenberth 2011). The period of enhanced hydrological activity in the 1970s corresponds to what McEwen (2006) terms a flood cluster period. Interestingly this period occurs just prior to a global shift in climate around 1976/1977, after which El Ninõ/La Ninã events increased/reduced in frequency; and the relationship between ENSO and rainfall in Thailand increased (Wang 1995; Trenberth and Hoar 1997, Singhrattna et al. 2005a,b).

The Ping river data suggest that ENSO events work together with tropical storms and/or monsoon irregularities to produce anomalous Q_{peak}, Q_{total} and Q_{min} (Table 2). The enhanced inverse relationship between ENSO activity and Thai rainfall should produce a weaker monsoon with less rainfall (and hence decreased runoff) in El Ninõ years; and the opposite effect should occur in La Ninã years (Singhrattna et al. 2005a,b). This general relationship is seen throughout the 89-year Q_{total} and Q_{wet} time series, and to a smaller extent, in the Q_{peak} time series (Figs. 2 and 3). However, closer inspection of the peak flow data and ENSO occurrences from 1980 onwards shows a more complicated relationship: e.g., El Ninõ years do not necessarily result in low Q_{peak} values; and vice versa for years with La Ninã events—even after 1980 when a stronger relationship between ENSO and Thai rainfall is expected. For example, there are several La Ninã years when above-average Q_{peak} values did not occur (e.g. 1988, 1998/99, 2007). Furthermore, above-average Q_{peak} was associated with El Ninõ years 1987 and 2006. Therefore, an unequivocal relationship between ENSO events and extreme high flows on the Ping River cannot be determined from the data examined. This could be due to the interactions between ENSO events, tropical storms and monsoon anomalies that result in anomalous flows experienced in this catchment, masking a clear relationship between ENSO events and discharge (extreme high events in particular). Our Q_{peak} dataset, however, shows increased variability in peak flows following the global climate shift suggested by climatologists (Wang 1995; Trenberth and Hoar 1997) (Fig. 4a).

The effect of ENSO and other climatic phenomena on low flows is unclear as there are years when low flows occurred regardless of the occurrence of an ENSO event (Figs. 2b, 3c and 5). The decrease in Q_{min} to below-averaged values from the mid-1950s occurred even when rainfall (RF_{annual}, RF_{wet} and even RF_{dry}) showed no sign of decrease. Low Q_{total} values (below -1σ) coincide with monsoon anomalies or El Ninõ events (Table 2). The influence of climatic variables on Q_{min} is however not so clear. Seven of the 14 years when Q_{min} was below -1σ occurred in years when a tropical storm and or monsoon anomaly occurred—the same phenomena often associated with high flows (Table 2). It is also likely that the low dry season flows after 1955 result, in part, from anthropogenic factors (e.g., dam management, land-cover change effects on water balance partitioning), rather than decreases in rainfall.

5.2 Anthropogenic Amplifiers and Modifiers

Major land-use changes occurred in the catchment during the 1950s and 1960s that were related to large-scale logging, growth of road networks, and development of highland agriculture (Fig. 6). Our land-cover change analysis for the period 1973 to 2005 verifies a decrease in forest cover (from 83 % percent to 67 %) and a 15 % increase in agricultural lands (Table 4). Together agriculture lands and barren lands make up about 32 % of the 2005 land cover, approximately double that from 1973. During this period many highland agriculture practices and production systems changed from predominantly subsistence

Table 4 Change in percentage land cover (1973–2005) for the upper Ping River catchment above the P1 station

	Water (%)	Forest (%)	Barren (%)	Cropland (%)
1973/1974	0.22	83.23	2.81	13.7
1988/89	0.53	77.00	2.90	19.6
2000	0.65	69.17	4.36	25.8
2005	0.74	67.17	3.64	28.5
Percentage change	0.52	−16.06	0.83	14.8

Based on analysis of Landsat imagery

farming to commercial farms, involving year-round multiple cropping systems, plantations, and greenhouses (Ziegler et al. 2009b). Some intensive agriculture practices require substantial irrigation in the dry season. The surfaces of these types of farming systems often have a higher propensity than original forest cover to generate surface runoff, perhaps to a degree that stream flow variables could be affected at some scales (Ziegler et al. 2009a). This is also true for road networks, which have expanded greatly in northern Thailand in the last several decades (Ziegler et al. 2001, 2004; Delang 2005).

Importantly, the 1970s to 1990s witnessed an increase in construction of dual purpose reservoirs for wet-season flood control and dry-season irrigation. Currently, there are three major structures controlling water flow above the Chiang Mai gauging station (Lebel et al. 2009): Mae Faek Weir (built in 1936); the Mae Taeng Weir (1973), and the Mae Ngad Dam (1985). The typical strategy for most dams is to reach storage capacity by the end of the rainy season in order to maximize water availability in the dry season. Water is also released year round to maintain base flows in the rivers downstream. Some of the decrease in low flows post-1960 (Fig. 2b) could be related to difficulties in managing these flows throughout the year as dry-season farming requires irrigation water and extensive extraction could contribute to reductions in dry season flows (Ziegler et al. 2009a). Also, there remains uncertainty about the potential for stream desiccation caused by large-scale tree plantations such as rubber and fruit trees (Ziegler et al. 2009a; Guardiola-Claramonte et al. 2008; 2010). Increasingly water is being extracted by individuals from tributary streams and lowland canal systems for domestic use and small-holder farming. Most extraction is not sanctioned, and therefore competes with water capture systems that supply rural as well as urban communities (Lebel et al. 2009).

Accompanying land-use changes in the rural Chiang Mai area is the expansion of Chiang Mai city into surrounding agricultural lands in more recent years. Greater Chiang Mai city area experienced a farm-to-city transition where built-up areas increased from 9 % to over 33 % between 1989 and 2009 (Sangawongse 2006; Lebel et al. 2009; Sangawongse et al. 2011). The effects of urbanization which include the construction of major ring roads and changes in the Ping river bank (narrowing and lining the river banks) and floodplains are blamed for causing more serious floods in recent years (Rigg and Ritchie 2002; Jompakdee 2004; Lebel et al 2009; EPSEA 2010).

The observed trends in stream flow variables in the Ping River data set do not correspond with any of these land-cover/land-use changes. For example, the Q_{peak} and Q_{total} time series show a general negative trend since 1973—opposite of what is [commonly] believed to result from wide-spread land-cover change and road expansion in Thailand (cf. Forsyth and Walker 2008). Furthermore, the low flows in the dry season show an increasing trend from the 1980s (Figs. 2 and 3a). In general, a 16 % change from forest to other land covers would not likely produce noticeable changes in stream flow variables in such a large catchment, where rainfall may occur in only a fraction of the area at any given time (cf. Bruijnzeel 2004). Two prior studies failed to find significant streamflow changes in large Thai rivers caused by deforestation (Alford 1992; Wilk et al. 2001). Expansion of the road network was also likely not substantial enough to produce seasonally high peak values, but it may have contributed to reduced low flows in the dry season in some sub-catchments (cf. Cuo et al. 2008).

Finally, anthropogenic causes of changes in river flows at the scale of the Ping River are difficult to confirm, partly because these activities affect river flows at different scales. Storage and release of reservoir water have a large-scale impact on downstream flows whereas changes in hydrological flow pathways associated with forest clearing (for agriculture purposes) have very localized impacts (e.g. Cuo et al. 2008) which may become masked at the scale of the upper Ping River system. In addition, the effects of anthropogenic changes may also be masked by unusual climate phenomena (see discussion in Section 5.1). The 2005 Q_{peak}, for example, corresponds to a 30-year flood, may have been influenced by anthropogenic activities, but these effects are overshadowed by the influence of heavy rainfall associated with the occurrence of several late-season typhoons (Wood and Ziegler 2008; Lim and Boochabun 2012).

6 Conclusions and Management Implications

We did not find significant changes in annual peak flows for the 6,355 km^2 Upper Ping River catchment over the 89-year study period (1921–2009), nor during a shorter period when wide-spread land-use change was occurring (1960 to 2009). There was however a tendency for annual peak discharge fluctuations to be more extreme from the late 1960s. Although extensive land-cover/land-use change has been a major environmental concern in Thailand since the middle of the last century, it is impossible to see any effects on high flows at this scale. We found that peak flow activity—and by extension, flooding—in the Ping River catchment is driven predominantly by climate phenomena, such as tropical storm activity, monsoon anomalies and ENSO events. These phenomena, which often act together in an anomalous year, tend to mask the possible influences of anthropogenic activity, including land-cover conversion, road-building, urbanization and, possibly, dam/reservoir management. Annual low flow value fluctuations however show a major decline beginning in the mid-1950s. While high flows are controlled largely by climate phenomena, low flows reflect the dominant impact of anthropogenic activities, including retention behind large dams, water extraction in tributary streams, and potentially water balance partitioning related to land-cover/land-use changes—although we were unable to verify the latter.

The lack of a statistical trend for annual and seasonal rainfall over the 89-year period, and a clear relationship with the time series patterns of runoff variables, highlight the fact that physical connections between climatic influences and hydrological behavior in such a large catchment are difficult to discern, especially when they occur together with anthropogenically-induced land-use changes. Lack of reliable data—particularly spatially-distributed rainfall data—hinders our ability to make accurate assessments especially since reliable long-term rainfall data were available from only one rainfall station. Although the relationships between tropical storm activity, monsoon anomalies and ENSO events are still uncertain (Chia and Ropelewski 2002; Webster et al. 2005; Wang et al. 2007; Grossmann and Morgan 2011), we were able to verify the important role of tropical storms in causing streamflow extremes. One future concern is that tropical storm activity may increase as a consequence of global warming and acceleration in the hydrological cycle. Wang et al. (2007) estimated that between 2 and 6 tropical storms will make landfall in Indochina in La Ninã years; but less than 2 in El Ninõ years. Future La Ninã events may therefore have a higher chance of contributing to more severe floods given the possibility of more tropical storms landing during these periods—perhaps on the order of scale as the 2011 Chao Phraya flood both for Chiang Mai city or cities downstream.

Given the uncertain relationship between high flows and climatic forcers, it may be wise to consider changing development trajectories to ones that will help reduce the impacts of future floods and droughts for the upper Ping River catchment as well as other catchments in Thailand. The continued expansion onto high-risk floodplain areas makes people and infrastructure more vulnerable to very large floods, especially since they often place their confidence in engineering works (EPSEA 2010; Ziegler et al. 2012a,b). Also of concern is the increasing use of dry-season water that must either be extracted from ground-water stores or sequestered in the wet-season in dual-purpose reservoirs. Improved weather forecasting (tropical storm monitoring, spatially-distributed rainfall monitoring) is needed to help manage water release to both maximize wet-season flood protection and dry-season water availability.

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