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Storage under the 2011 Chao Phraya River flood: An interpretation of watershed-scale storage changes at two neighboring mountainous watersheds in northern Thailand

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Abstract:

The present study attempted estimations of watershed-scale storage changes at two mountainous watersheds in northern Thailand to understand the behaviors of watershed-scale storage under the 2011 Chao Phraya River flood. For this purpose, we applied a methodology that separates an hourly hydrograph into several discharge sub-components, and formulized watershed-scale storage-discharge relationships. The results showed that (1) this methodology was applicable to sub-tropic watersheds, (2) there were five different discharge sub-components, that correspond to the number of dominant rainfall-runoff processes, in two mountainous watersheds in northern Thailand, (3) the peak total storage in 2011 was estimated to occur in October because of strongly seasonal slower discharge sub-components, whereas the maximum total discharge was observed in June, (4) the sum of watershed-scale maximum storages of all the discharge sub-components in the upper Yom and Nan River watersheds were respectively estimated to be 135 mm and 405 mm, and the difference might be explained by the existence of the active fault running north-south in the upper Nan River watershed, and (5) the estimated storage with the recession time constants of 111 h at the beginnings of rainy seasons could explain the risk of slope failure occurrences within a watershed.

KEYWORDS hydrograph separation; recession analysis; flood; Yom River; Nan River; slope failure

INTRODUCTION

The 2011 flood in the Chao Phraya River watershed received international attention and several reports have been published from a hydrological viewpoint (e.g. Komori *et al.*, 2012; Ziegler *et al.*, 2012), but none of the reports directly interpreted the flood in terms of watershed-scale storage behaviors. Hence, we have no effective clues to answer the following questions: (1) *how did the watershed-scale storages change?*, (2) *how much water was stored in the watershed?*, (3) *what was special about the watershed-scale storage in 2011?*, *and* (4) *can the watershed-scale storage explain the slope failures that occurred in 2011?*.

Why haven't the watershed-scale storage changes been directly interpreted yet? It would be because storage changes are difficult to observe or estimate at a watershed-scale. In the watershed-scale water balance equation,

$$P = ET + Q + \mathrm{d}S/\mathrm{d}t \qquad (1)$$

where P, ET, Q, $\mathrm{d}S/\mathrm{d}t$ are respectively intensities of precipitation, evapotranspiration, discharge, and storage change, P and Q are observed in many places of the Chao Phraya River watersheds, we can calculate $\mathrm{d}S/\mathrm{d}t$ by estimating ET. Yet, the estimation of ET at shorter time scales is still difficult at a watershed-scale, because we cannot verify the magnitude of ET at the spatial scale from observational data such as in Kim *et al.* (2013). Conversely, storage change $\mathrm{d}S/\mathrm{d}t$ is also difficult to observe or estimate at a watershed scale. Therefore, strictly speaking, both ET and $\mathrm{d}S/\mathrm{d}t$ remain as unknowns and hence understanding of the watershed-scale water balance is still one of the fundamental questions of watershed-hydrology even today (e.g. Reggiani *et al.*, 2000; Sivapalan, 2003; Beven, 2006; Yokoo *et al.*, 2008).

Toward a solution for the above problem, a traditional but novel approach was introduced by Kirchner (2009) who directly related $\mathrm{d}S/\mathrm{d}t$ and Q by assuming ET and P are both zero in rainless night-time periods and formulizing the storage-discharge relationship as Equation (2).

$$\mathrm{d}Q/\mathrm{d}S = -(\mathrm{d}Q/\mathrm{d}t)/Q \qquad (2)$$

This equation means a storage-discharge relationship can be directly derived from the scatter diagram between $-\mathrm{d}Q/\mathrm{d}t$ and Q. His work received considerable attention and was followed by many studies related to applications and improvements (e.g. Teuling *et al.*, 2010; Birkel *et al.*, 2011; Krakauer and Temimi, 2011; McMillan *et al.*, 2011; Sayama *et al.*, 2011). The first author also applied this methodology and found that a single storage-discharge formulization is insufficient in Japanese temperate watersheds (Yokoo *et al.*, 2012) and suggested in Kobayashi and Yokoo (2013) incorporating the hydrograph separations method by Hino and Hasebe (1984) with the method of Kirchner (2009). Based only on observed data, the method of Kobayashi and Yokoo (2013) allows identification of dominant rainfall-

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runoff processes by separating total discharge into its sub-components as well as their corresponding storage behaviors, although the applicable watershed size of the method of Kirchner (2009) remained unexplored. Yet, by taking this approach, we do not need to stick only to "hydrograph matching" but rather we can monitor the behaviors of storage as the "soft-data" (Seibert and McDonnell, 2002) for reducing uncertainties of model structures and parameters that are often questioned even in current distributed hydrological modeling (e.g. Yokoo and Kazama, 2012; Mateo *et al.*, 2013). Hence it is worth testing the method of Kobayashi and Yokoo (2013) to interpret the watershed-scale storage behaviors under the 2011 Chao Phraya River flood.

The present study attempts the first application of the method of Kobayashi and Yokoo (2013) in two mountainous watersheds in northern Thailand for interpreting the 2011 Chao Phraya River flood in terms of watershed-scale storage behaviors. Then we investigated the relationship between watershed-scale "dynamic" storage that eventually becomes discharge and the occurrence of slope failure in 2011 to see if our estimates of watershed-scale storage could explain the occurrences of slope failures, where applicable watershed-size of this method remains unexplored to be reported in a separate paper.

METHOD

Study area and data

The present study selected the watersheds of "Y.20" and "N.64" discharge monitoring stations (hereafter Y.20 and N.64 watersheds) that are respectively located at the upper most streams of the Yom and the Nan Rivers in northern Thailand as shown in Figure 1. Both rivers overflowed at their lower flood plains because of the repeated heavy rainfall in the Chao Phraya River watershed in 2011 (Komori *et al.*, 2012). The attributes of the two basins are summarized in Table I.

The authors prepared discharge data of the Y.20 and the N.64 watersheds observed and quality-checked by the Royal Irrigation Department (RID), Thailand for the local hydrological years of 2009–2011 starting from April. We also prepared watershed-scale hourly precipitation data for the same period from the Global Satellite Mapping of Precipitation (Kubota *et al.*, 2007) to specify rainless periods within the watersheds as well as ground-based daily precipitation data measured by the RID and the Thai Meteorological Department (TMD) to estimate watershed-average precipitation for confirming the annual water balance in the watersheds. We used 2 daily precipitation monitoring stations (Song and Chiang Muan) in the Y.20 watershed and 4 stations (Pua, Thung Chang, Tha Wang Pha, Chiang Klang) in the N.64 watershed.

Hydrograph separation

Kobayashi and Yokoo (2013) separated hourly hydrographs by the filter-separation auto-regressive method of Hino and Hasebe (1984), after converting the discharge unit from m^3/s to mm/h by dividing with the watershed area in Table I. Details of the method are described in Hino and Hasebe (1984), and here we briefly explain the outline of the method.

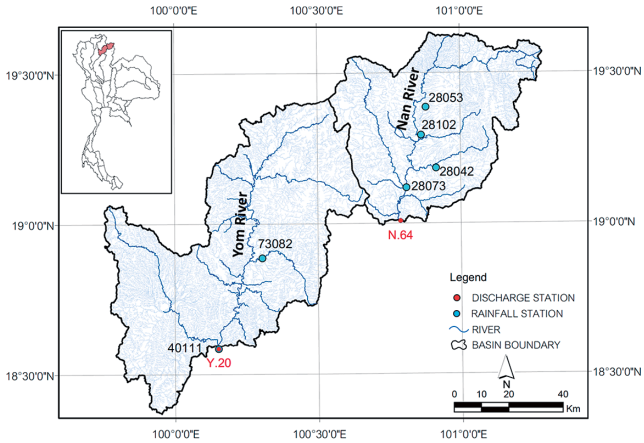


Figure 1. Geographic locations of the Y.20 and the N.64 watersheds, where "40111", "73082", "28073", "28042", "28102", and "28053" are the code numbers of rainfall monitoring stations

Table I. Attributes of the Y.20 and the N.64 watersheds derived from a digital elevation map in Thailand. The length of main river channel and total length of channel network are obtained from the vector data of the channel in the geographic data in Thailand. The drainage density was calculated as the total length of channel network divided by the drainage area

	Y.20	N.64
Drainage area (km^2)	5,394	3,476
Length of main river channel (km)	172	169
Total length of channel network (km)	1,088	756
Mean gradient from river profile	0.0138	0.0110
Maximum elevation (m)	1,629	1,676
Minimum elevation (m)	186	219
Mean elevation (m)	538	659
Standard deviation of elevation (m)	212	326
Mean width of watershed (km)	31.4	20.6
Shape factor of watershed	0.183	0.122
Drainage density (km^{-1})	0.2017	0.2175

Firstly this method finds a characteristic recession period and applies an exponential function to obtain its recession exponents. With the exponent, Hino and Hasebe (1984) formulize a numerical filter to separate into faster and slower discharge components. If we repeat this separation, we can decompose a hydrograph into several discharge sub-components. From our experience, we defined the number of sub-components to be around 3 to 5, coinciding with the number of tanks of the Tank Model applied in many watersheds (Sugawara, 1995; Yokoo *et al.*, 2001). The parameters obtained from hydrograph separations are summarized in Table II.

To reduce the effect of initial conditions on the hydrograph separation method, we removed the data of the 2009 hydrological year and used those of 2010–2011 for the following analysis.

Estimations of watershed-scale "dynamic storage"

Kobayashi and Yokoo (2013) estimated watershed-scale

Table II. Parameters used in hydrograph separations and storage-discharge relationships. The lower sub-scripts correspond to slower components. "A" with subscripts are adjusting parameters in the numerical filter in Kobayashi and Yokoo (2013)

Method	Parameters	Y.20	N.64
Hydrograph separation	T_{c1} (h) (= $1/\beta_1$)	951	2840
	T_{c2} (h) (= $1/\beta_2$)	155	637
	T_{c3} (h) (= $1/\beta_3$)	61.2	111
	T_{c4} (h) (= $1/\beta_4$)	12.3	26.4
	T_{c5} (h) (= $1/\beta_5$)	—	—
	A_1	0.20	0.10
	A_2	0.20	0.18
	A_3	0.20	0.30
	A_4	0.50	0.70
	A_5	—	—
Storage-discharge formulization	a_1	0.0000623	0.000289
	a_2	0.00177	0.00133
	a_3	0.00641	0.00370
	a_4	0.0180	0.00756
	a_5	0.0718	0.0430
	b_1	0.447	1.04
	b_2	0.961	1.11
	b_3	1.03	1.11
	b_4	1.03	0.984
	b_5	1.12	1.08

"dynamic" storage, which eventually becomes discharge, based on Kirchner (2009) which assumes the following relationship holds in rainless nighttime recession periods where both P and ET are negligible.

$$\frac{\mathrm{d}Q}{\mathrm{d}S} = \frac{\mathrm{d}Q/\mathrm{d}t}{P - ET - Q} \cong \left.\frac{-\mathrm{d}Q/\mathrm{d}t}{Q}\right|_{P \ll Q \,\&\, ET \ll Q} \tag{3}$$

Then Kirchner (2009) demonstrated that we can define a watershed-scale storage-discharge relationship by introducing the power-law recession model in Brutsaert and Nieber (1977) as an idealized approximation to obtain Equation (4).

$$\frac{\mathrm{d}Q}{\mathrm{d}S} \cong aQ^{b-1}\Big|_{P \ll Q \,\&\, ET \ll Q} \tag{4}$$

We can easily obtain the parameters a and b by finding a best-fit power-law function. Finally, Kirchner (2009) integrated Equation (4) to obtain the storage-discharge relationship in Equation (5),

$$S - S_0 = \frac{1}{a} \cdot \frac{1}{2-b} Q^{2-b} \tag{5}$$

where S_0 is an integral constant that is equivalent to the storage level at which discharge becomes zero. Although S_0 remains unknown, we can estimate "dynamic" storage S-S_0 that eventually becomes discharge.

Kobayashi and Yokoo (2013) applied this method for relating all the discharge sub-components with their corresponding "dynamic" storages by using the data in rainless nighttime recession periods for all the components. Yet, we applied this method only for the fastest discharge sub-component with shortest recession time constant by using the data in "rainless" nighttime recession periods and we used the data in nighttime recession periods for other discharge sub-components because slower sub-components would have less effect on precipitation in larger watersheds. The nighttime in the present study was defined to be 19:00 to 6:00 (of the next day) at the local time. Note that the estimated "dynamic" storages are not interconnected in series or parallel with each other, and hence our storage-discharge relationships are different from these of multiple reservoir models such as the Tank Model (Sugawara, 1995; Yokoo *et al.*, 2001). The relationship between our method and the multiple reservoir models is currently being investigated for discussion in a separate paper.

RESULTS

Hydrograph separation and estimation of watershed-scale dynamic storage

Figure 2 shows the results of hydrograph separations in panels (a) and (b) and estimations of watershed-scale storages in panels (c) and (d) in the Y.20 and N.64 watersheds. The parameters used are summarized in Table II.

As in the Figure 2 (a) and (b), the hourly discharge data of the Y.20 and the N.64 watersheds were both separated into 5 sub-components. We can see that the discharge of the N.64 watershed was higher than the neighboring Y.20 watershed consistently both in 2010 and 2011. As reported by Komori *et al.* (2012), the higher discharge (> 0.1 mm/h) period in 2011 is longer than that of 2010 where the instantaneous discharge intensities were almost the same in both years. Also, we can see that the intensities of slower discharge sub-components named Q_1, Q_2, and Q_3 were about the same between 2010 and 2011 in the two watersheds and faster discharge sub-components named Q_4 and Q_5 were higher in 2011 indicating the relative dominance of faster discharge sub-components.

The panels (c) and (d) show the watershed-scale "dynamic" storage estimated by applying Equation (5) for all the discharge sub-components and their storage changes in the Y.20 and the N.64 watersheds. The estimated dynamic storages are dominated by the storages of the slower discharge sub-components of Q_1 and Q_2, indicating the dominance of seasonality of slower discharge sub-components as the result of strongly seasonal rainfall. We can see that the peak of the total dynamic storage in 2011 is about twice that of 2010 in the Y.20 watershed, whereas they are almost the same in the N.64 watershed. This difference indicates the different roles of storages of faster discharge sub-components such as Q_3, Q_4, and Q_5 in the two watersheds. In the Y.20 watershed, the fractions of the storages of faster discharge sub-components are small where increased rainfall in 2011 caused an increase of storages of slower discharge sub-components. In the N.64 watershed, the storage fractions of faster discharge sub-components are higher. This result in the N.64 watershed potentially causes higher total storage in 2011 than 2010, but the peak storages for the faster discharge sub-components in 2011 occurred on June 27 whereas the storages of the slower discharge sub-components were at the beginning of seasonal increases.

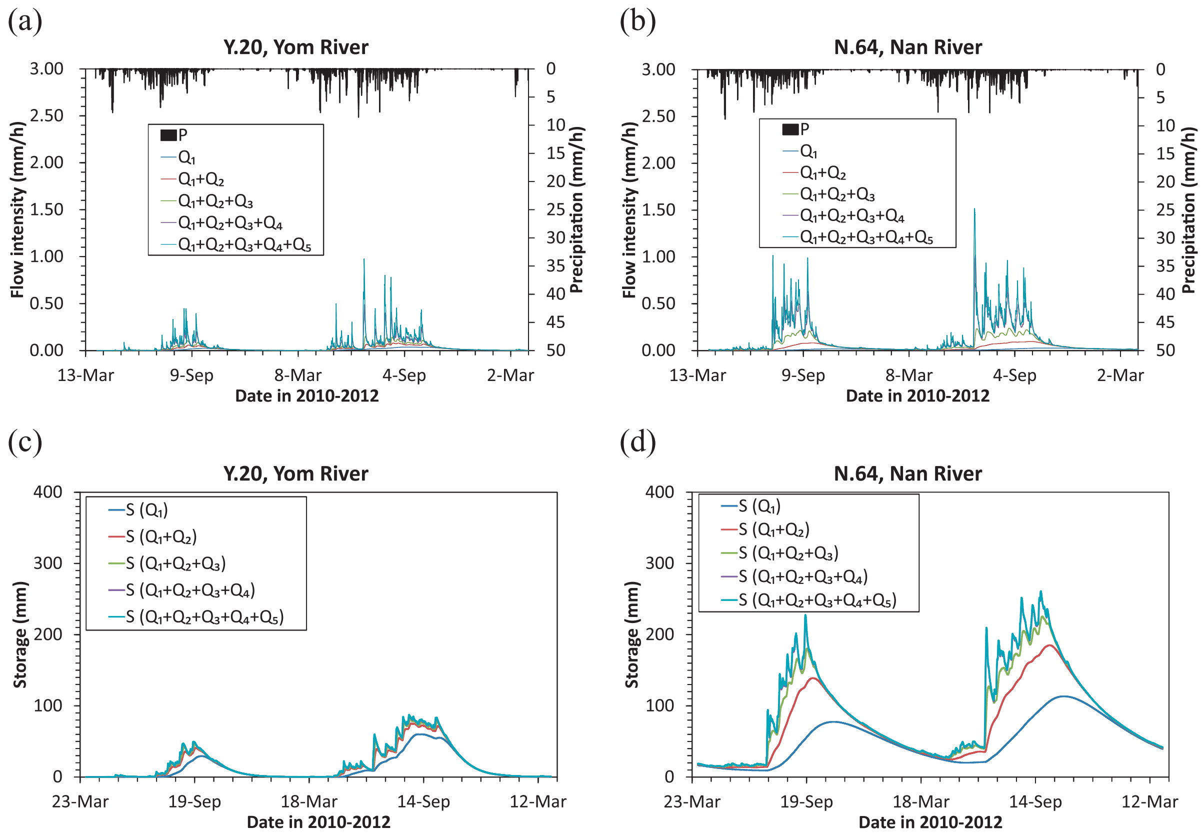


Figure 2. Results of hydrograph separations and estimated watershed-scale storage changes. The panels (a) and (b) are separated hydrographs in the Y.20 and N.64 watersheds. The panels (c) and (d) are estimated watershed-scale storage changes in the two watersheds

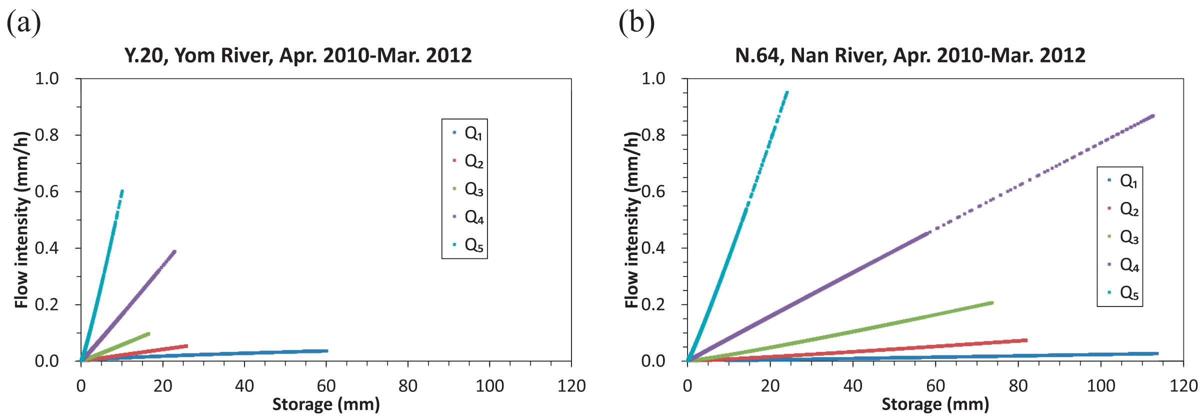


Figure 3. Relationship between watershed-scale storage and discharge for different discharge sub-components in (a) the Y.20 watershed and (b) the N.64 watershed

Hence the peak storage of the faster discharge sub-components in June 27 did not become the highest total storage over an annual time scale. Namely, the difference of the peak timing between faster and slower discharge sub-components should have made the peaks of total storages similar between 2010 and 2011.

Storage-discharge relationships

Figure 3 shows the relationships between watershed-scale storages and discharge for different discharge sub-components in the Y.20 (a) and the N.64 (b) watersheds. Both of the panels (a) and (b) clearly show that the relationship between watershed-storage and discharge for all the discharge sub-components were almost linear in the ranges of watershed-scale storage and discharge. The discharges of 2011 in the two watersheds are of a severe flood year and they should cover their almost full ranges, and hence we expect that the relationship between storage and discharge in the two basins could be generally modeled by a combination of multiple linear reservoirs as pointed

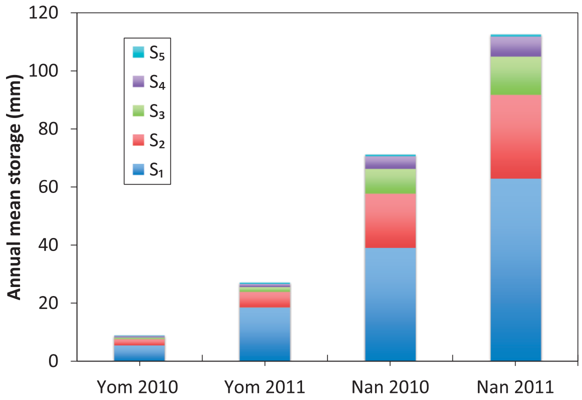


Figure 4. Annual mean storages of the Y.20 (Yom) and the N.64 (Nan) watersheds in 2010 and 2011. "S_1" to "S_5" indicate constituting storage levels, where smaller numbers after "S" mean storages of slower discharge sub-components

out by Hino and Hasebe (1984).

Under the severe flood of Thailand in 2011, what was the sum of instantaneous maximum storages of discharge sub-components in our estimates? It was about 135 mm in the Y.20 watershed and 405 mm in the N.64 watershed. These estimates are comparable to the constant storage capacity of 450 mm used in a distributed hydrological model in the same area of Thailand (Mateo *et al.*, 2013; Hanasaki, 2013). Note that we are discussing about the "dynamic" storage that was assumed to be zero when all the discharge sub-components became zero, hence the residual "static" storage are excluded from our discussion.

Annual mean watershed-scale storages

Figure 4 shows the annual mean watershed-scale storages in the Y.20 and the N.64 watersheds for 2010 and 2011. Surprisingly, the annual mean watershed-scale storage in the N.64 watershed is about 6 times higher than that of the Y.20 watershed in 2010. In the case of 2011 with the severe flood, the annual mean watershed-scale storage of the N.64 watershed was about 5 times higher than that of the Y.20 watershed.

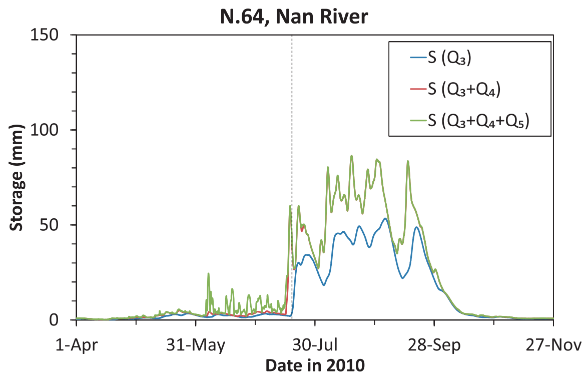
These estimates suggest there should be significant effects of different physiographic characteristics between the two watersheds, although they are neighboring each other as in Figure 1. Based on the ground-based precipitation monitoring data, the arithmetic means of annual precipitations at Song (station number: 40111) and Chiang Muan (station number: 73082) in the Y.20 watershed were 1,252 mm/y in 2010 and 1,613 mm/y in 2011. In the N.64 watershed, those at Pua (station number: 28042), Thung Chang (station number: 28053), Tha Wang Pha (station number: 28073), and Chiang Klang (station number: 28102) were 1,400 mm/y in 2010 and 1699 mm/y in 2011. Hence the annual precipitations in the Y.20 and the N.64 watersheds would be almost the same in both 2010 and 2011, indicating that the difference in precipitation is not the cause of the difference in the storage level in the two watersheds. There are no remarkable differences in topography, vegetation cover, soil type, and land use between the two watersheds. Yet, according to Soralump (2013), there is an active fault line from the northern end to southern end of the N.64 watershed that has fractured zones along the fault line, which could result in higher infiltrations, storage levels and drainability in the N.64 watershed. Currently, we guess that the existence of the active fault line is the most possible reason for the higher storage level in the N.64 watershed.

Relationship between near surface storage and occurrence of slope failure

If we can assume that the watershed-scale dynamic storages of the faster discharge sub-components Q_3, Q_4, and Q_5 are near the ground surface of a watershed, it may be possible to discuss the relationship between the near surface storage changes and occurrences of slope failure within a watershed.

Figure 5 shows the total dynamic storages of the faster discharge sub-components Q_3, Q_4, and Q_5 in the N.64 watershed that experienced slope failure in 2010 (a) and 2011 (b), where no slope failure occurred in the Y.20 watershed in 2010 and 2011. The storages of the three discharge sub-components of Q_3, Q_4, and Q_5 are selected because the recession time constants were smaller than or equal to 111 hours in the N.64 watersheds as in Table II, which seemed to be reasonable for discussing the occurrences of slope failure caused by the increase of near

(a)



(b)

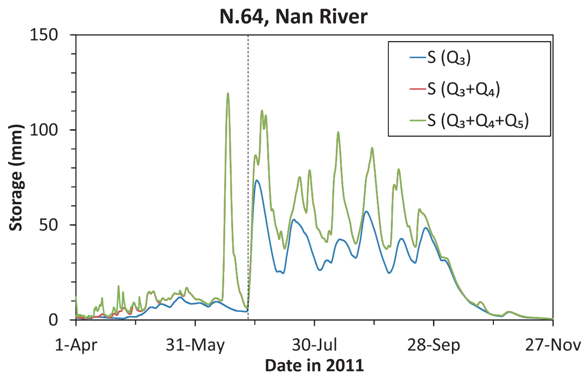


Figure 5. Cumulative storage levels for the faster discharge sub-components of Q_3, Q_4, and Q_5 of the N.64 watershed in (a) 2010 and (b) 2011. The vertical dash line segments indicate the dates of slope failures occurred in 2010 and 2011 in the N.64 watershed, where no slope failures occurred in the Y.20 watershed

surface dynamic storages.

The vertical dash line segments in Figure 5 show the date of slope failure occurrences within the N.64 watershed in 2010 and 2011. According to the Department of Mineral Resources (2011), slope failure occurred on July 18, 2010 at Ban Nam Ki M.5, Ban Sanchareon M.6, Tambon Phathong, and Amphoe Tha Wang Pha, because of the tropical storm Conson. On June 26, 2011, tropical storm Haima caused slope failure at Amphoe Bo Klua, Thung Chang, Pua, and Tha Wang Pha.

From Figure 5, we can confirm that both of the slope failures in 2010 and 2011 occurred between the first peak total storages of Q_3, Q_4, and Q_5 and the first seasonal increases of watershed-scale storage of Q_3 at the beginnings of the rainy seasons in June and July. In the case of 2010, a slope failure occurred on July 18 when the sum of storages of Q_4 and Q_5 was just after its peak on July 17 at 59 mm and the storage of Q_3 started to increase toward its first seasonal peak at about 30 mm on July 21. In the case of June 26 of 2011, slope failure occurred when the sum of storages of Q_4 and Q_5 was just after its peak on June 16 at 113 mm and the storage of Q_3 started to increase toward its peak at 74 mm on June 30. In other words, risk of slope failure occurrences within the N.64 watershed increases when storage of Q_3 starts to increase at the beginning of a rainy season.

These processes can be interpreted as the dried N.64 watershed at the end of the dry season in March absorbing rainfall in April and May to increase the storages of Q_4 and Q_5 to generate higher *ET* and lower *Q*. Because of the strong storms in June or July, the storage of Q_3 starts to increase because of the higher near surface storages of Q_4 and Q_5 compared with the dry season, which increases the risk of slope failure occurrences within the watershed.

These results suggest that the watershed-scale storages estimated by our method would be potentially useful for estimating occurrences of slope failure within a watershed. Namely, we might be able to realize the risk of slope failures by monitoring the storage of Q_3 in the case of the N.64 watershed.

DISCUSSIONS

What is novel and what became possible?

The present study estimated the watershed-scale storage changes in the Y.20 and the N.64 watersheds in northern Thailand as the first attempt to apply the method of Kobayashi and Yokoo (2013) in the sub-tropic watersheds. As in Supplement Figure S1 and S2, we faced a problem of discharge data limitations of the two watersheds. The watersheds are under a strongly seasonal climate, which results in a cycle of seasonal increase and decrease in the slowest discharge sub-component at the beginning and the end of a rainy season, respectively. Hence, the relationship between dQ/dt and Q for the slowest discharge sub-component became very sparse. This must be a characteristic problem of a watershed with a strongly seasonal climate. Because there was no such problem in Japanese watersheds (Kobayashi and Yokoo, 2013) with weaker climatic seasonality than the Y.20 and the N.64 watersheds, it is a novel finding of the present study.

Nonetheless, by using the method of Kobayashi and Yokoo (2013), we could discuss the watershed-scale storages for different discharge sub-components during 2010–2011. As the results, we obtained the following findings.

(1) The "dynamic" storage of the N.64 watershed was about 5 to 6 times higher than the Y.20 watershed, which had not previously been reported in relation to the flooding in 2011.
(2) The instantaneous watershed-scale total storage height of 2011 in the Y.20 watershed became higher than 2010, whereas they are almost the same in the N.64 watershed. Also, the peak storages are estimated to have occurred in October of 2010 and 2011 because of the strong seasonality of the watershed storages of the slower discharge sub-components. The characteristic behavior of the watershed-scale storage in 2011 was not the height but the length of high-storage period compared to that of 2010, which coincides with the report on discharge behaviors by Komori *et al.* (2012).
(3) Under the assumption that the watershed-scale storages for the faster discharge sub-components correspond to those of the near surface storages, we found a possibility that we might be able to estimate the occurrences of slope failures within a watershed by monitoring the watershed-scale storage heights for the discharge sub-components with recession time constants shorter than or equal to 111 hours. Although such an idea is similar to the "Soil Water Index" introduced by Okada *et al.* (2001), we think that our advantage lies with our deterministic methodology for the model structure based only on the observed data and the easy parameter identifications.

Problems of proposed methodology and possible future directions

The present study was conducted assuming that the methodology of Kobayashi and Yokoo (2013) is applicable at the mountainous watersheds in northern Thailand to explore its applicability in sub-tropical watersheds. As a result, we faced three unique problems as follows.

(1) We could use very limited data for the slowest discharge sub-component because of the strongly seasonal discharge in the target watersheds, and the scatter diagram between dQ/dt and Q became very sparse and the representativeness of the power-law regression curve became considerably low compared to the first attempt by Kirchner (2009) in the UK and even to the application by Kobayashi and Yokoo (2013) in Japan.
(2) Not only the sparseness of the scatter diagrams but also the higher deviations of the scatters from the power-law regression curves in the scatter diagram between dQ/dt and Q compared to the corresponding diagram in Kirchner (2009) were another unique methodological problem in the study watersheds, which was also found by Kobayashi and Yokoo (2013) in Japan. The reason for this problem would come from the relatively complex hydrological processes in the watersheds in northern Thailand and Japan (Kobayashi and Yokoo, 2013). The authors think that the examples introduced by Kirchner (2009) in the UK were rather a rare case exhibiting relatively simple hydrological processes where the scatter diagram between dQ/dt and Q became densely

concentrated along the power-law regression curves where no hydrograph separation were necessary.

(3) Although the study watersheds are under sub-tropic climate where night-time evapotranspiration could take place, we assumed that evapotranspiration is zero from 19:00 to 6:00 of the next day in the local time. According to the simulation by Kim *et al.* (2005) in Thailand, the night-time evapotranspiration is not zero. If this is the reality, we should carefully remove such data from the scatter diagram between dQ/dt and Q. Now we are planning to revisit the data with the monitoring results in Thailand accessible through the "Agro-Meteorological Forecaster" maintained by Dr. Wonsik Kim at (http://matthew.niaes.affrc.go.jp/~wonsik/).

By solving the above problems, the authors will make this method more reliable in sub-tropic watersheds as a useful tool for the identification of dominant rainfall-runoff processes and model parameters. We could not comment on the applicable watershed-size for our methodology, however this will be discussed in a separate paper.

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SUPPLEMENTS

Figure S1. Relationships between dQ/dt and Q for Q_1 to Q_5 in the Y.20 watershed

Figure S2. Relationships between dQ/dt and Q for Q_1 to Q_5 in the N.64 watershed

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