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**UNCERTAINTY OF RUNOFF
ASSOCIATED WITH UNCERTAINTIES OF
WATER HOLDING CAPACITY AND
RAINFALL DISTRIBUTION IN
MOUNTAINOUS CATCHMENTS**

By

Intan Supraba

A doctoral dissertation submitted in partial fulfillment of the requirements for
the degree of Doctor of Philosophy in Engineering

Examination Committee:	Associate Professor Tomohito J. Yamada
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	Professor Yasuyuki Shimizu
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September 2015**

**UNCERTAINTY OF RUNOFF ASSOCIATED
WITH UNCERTAINTIES OF WATER HOLDING
CAPACITY AND RAINFALL DISTRIBUTION IN
MOUNTAINOUS CATCHMENTS**

(山地流域における保水能と降水分布の不確実性
がもたらす流出量の不確実性)



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A dissertation submitted to the Graduate School of Engineering of the Hokkaido
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Dedicated to my beloved parents:
Bapak Mochammad Joko Santosa & Ibu Twidayati

ABSTRACT

Simulating runoff with higher accuracy to reduce the flood risk is one of the research targets in hydrology. The runoff simulation in previous studies can be classified into two types of hydrological models such as lumped and distributed types. However, these hydrological models simulate runoff deterministically, when actually the timing and amount of peak runoff is sensitive to rainfall distribution, both in temporally and spatially due to the non-linear characteristics of rainfall-runoff processes. A previous study investigated the non-linearity of runoff phenomena in mountainous catchments, and proposed water holding capacity distribution theory based on total rainfall-total rainfall loss relationship to estimate effective rainfall intensity as the input data to simulate runoff. The relationship between total rainfall and total rainfall loss is well fitted using the tanh function fitting curve.

This study uses hourly rainfall and hourly runoff data obtained from the Ministry of Land, Infrastructure, Transportation, and Tourisms (MLIT), Japan database during summer and autumn (June-October) at least for 10 years (2002-2011). After checking the data quality among 106 catchments that available in the database, only 36 catchments have the continuous data set that fulfill the target period in this study. Thus, total rainfall-total rainfall loss relationship is applied to those 36 catchments, and results indicate that those 36 catchments can be classified into 2 groups i.e. 23 catchments having a constant-stage tanh-type curve, and 13 catchments having a non-constant-stage tanh-type curve. Based on the physical interpretations given before to the linear and constant parts of the tanh curves, catchments having a constant-stage tanh-type curve are characterized by a constant stage after the linear stage due to some heavy rainfall events that have small total rainfall loss, which catchments having a constant-stage tanh-type curve demonstrate saturation conditions. Thus, the effective rainfall intensity for simulating runoff only can be estimated for those 23 catchments having a constant-stage tanh-type curve.

From the obtained results, this study found that runoff parameter a in the total rainfall and total rainfall loss relationship represents the height of tanh curve that can be used to estimate the potential catchment storage for catchments having a constant-stage tanh-type curve. Thus, runoff parameter a is an important parameter that can be used to estimate the capacity of a catchment to hold or to store water during a rainfall event. By knowing the capacity of a catchment, the amount of rain water that become direct runoff that causes flooding can be estimated. However, the plotting result of total rainfall-total rainfall loss relationship show that similar values of total rainfall occurred in two different rainfall events have different values of total rainfall loss. This difference indicates the effect of initial soil moisture condition. The rainfall event that having bigger total rainfall loss means the respective rainfall event occurred when the catchment was initially dry, and on contrary, the rainfall event that having smaller total rainfall loss means the respective rainfall occurred when the catchment was initially wet.

The plotting result of total rainfall-total rainfall loss relationship also show the standard deviation values (1σ) of runoff parameters a and b . Thus, this study interpret the value of 1σ of runoff parameter

a as the initial water amount that reflects the initial soil moisture condition, and is used as the parameter to explain about the uncertainty associated with water holding capacity.

As many catchments in Japan have never experienced the saturation condition, the runoff parameters in the total rainfall-total rainfall loss relationship for estimating effective rainfall intensity cannot be obtained. Therefore, this study proposed two methods, namely localized gradient method and inverse method to estimate those runoff parameters. Then the estimated runoff parameters are validated by comparing the time to peak and peak runoff simulated using the estimated runoff parameters, and that of based on observation data.

The rainfall data by MLIT used in this study are measured by rain gauges that cover the whole Japan. However, actually rainfall as the input data to simulate runoff contains uncertainty. Rain gauges measure the rainfall intensity near to the land surface, but there is limitation on their spatial representativeness due to the location and density of rain gauges. The measured amounts are influenced by several factors such as wind, snowfalls, station relocation, and change of the sensors. A previous study concluded that gauges based rainfall intensity measurements can be biased by factors like wind and evaporation in the range of 10-20%.

Japan Meteorological Agency (JMA) estimates rainfall intensity by using C-band radar and X-band radar that having high spatial and temporal resolution over extended areas. However, radar also has uncertainty due to several factors such as hardware calibration, mountain blockage, and anomalous propagation. Thus, whether measured directly by rain gauges or indirectly by remote sensing techniques, all rainfall intensity measurement contain uncertainty.

The theory about uncertainty is analogous to the random term in Brownian motion. The first theory of Brownian motion is in consequence of the role of Gaussian variables in probability. The stochastic force and derivation of Einstein's theory of Brownian motion from Newton's second law was introduced by Paul Langevin. A recent study analyzed the uncertainty of peak runoff height using the stochastic differential equation (sde) method by analyzing the uncertainty of rainfall distribution where the probability of runoff height can be derived from the Fokker-Planck equation. Results showed that 10% uncertainty of rainfall distribution contributes to the uncertainty of peak runoff height.

In this study, the uncertainty of peak runoff height is investigated by considering two independent uncertainties i.e. uncertainty associated with water holding capacity, and uncertainty associated with rainfall distribution. Two different methods, ensemble method and sde method, are proposed to quantify the uncertainty of peak runoff height associated with those two uncertainties. Results show that the peak runoff height uncertainty increase with the increment of uncertainty associated with rainfall distribution, and uncertainty of water holding capacity needs to be included in the quantification of the uncertainty of peak runoff height. By utilizing the results of uncertainty of peak runoff height, the main objective of this study is to quantify the uncertainty of peak runoff associated with those two independent uncertainties. Results show that the uncertainty of peak runoff associated with water holding capacity is more dominant when the uncertainty of rainfall distribution is 10%, and it is less dominant when the uncertainty of rainfall distribution is 20%.

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TABLE OF CONTENTS

ABSTRACT	iv
ACKNOWLEDGMENT.....	vi
TABLE OF CONTENTS.....	vii
LIST OF FIGURES.....	ix
LIST OF TABLES.....	xi
Chapter 1. INTRODUCTION.....	1
1.1 RESEARCH BACKGROUND.....	2
1.2 RESEARCH OBJECTIVE.....	5
Chapter 2. METHODOLOGY.....	6
2.1 DATA.....	7
2.2 STREAMFLOW HYDROGRAPH SEPARATION.....	9
2.3 LUMPED MODEL.....	10
Chapter 3. POTENTIAL CATCHMENT STORAGE ESTIMATION	20
3.1 RELATIONSHIP BETWEEN TOTAL RAINFALL AND TOTAL RAINFALL LOSS.....	21
3.2 STANDARD DEVIATION OF PARAMETER a	28
3.3 SUMMARY.....	30
Chapter 4. PROPOSED ESTIMATION METHOD OF POTENTIAL CATCHMENT STORAGE FOR CATCHMENTS THAT NEVER EXPERIENCE SATURATION CONDITIONS.....	31
4.1 ESTIMATION OF RUNOFF PARAMETERS (a and b) AT THE CONSTANT STAGE	32
4.1.1 LOCALIZED GRADIENT METHOD.....	33
4.1.2 INVERSE METHOD	34
4.2 CASE STUDY.....	35
4.3 SUMMARY	38
Chapter 5. THE MINIMUM TOTAL RAINFALL REQUIRED TO GENERATE DIRECT RUNOFF.....	39
5.1 THRESHOLD OF MINIMUM TOTAL RAINFALL REQUIRED TO GENERATE DIRECT RUNOFF	40
5.2 VARIATION OF POTENTIAL CATCHMENT STORAGE	41
5.2.1 CATCHMENT MORPHOMETRIC PARAMETERS	41
5.2.2 CATCHMENT CHARACTERISTICS	44
5.3 SUMMARY.....	58

Chapter 6. UNCERTAINTY OF PEAK RUNOFF.....	59
6.1 HISTORY OF STOCHASTIC DIFFERENTIAL EQUATION	60
6.2 RELATIONSHIP BETWEEN ITO STOCHASTIC DIFFERENTIAL EQUATION AND FOKKER-PLANCK EQUATION.....	62
6.3 PROPOSED METHODS TO QUANTIFY UNCERTAINTY OF PEAK RUNOFF HEIGHT	69
6.3.1 ENSEMBLE METHOD	69
6.3.2 STOCHASTIC DIFFERENTIAL EQUATION METHOD	69
6.4 CASE STUDY	72
6.4.1 ENSEMBLE METHOD	72
6.4.2 STOCHASTIC DIFFERENTIAL EQUATION METHOD	75
6.5 SUMMARY.....	83
Chapter 7. CONCLUSIONS	84
7.1 CONCLUSIONS.....	85
REFERENCES.....	87
APPENDICES.....	92
RESUME	111
LIST OF PUBLICATIONS	113

LIST OF FIGURES

Figure 2.1. Geographic distribution of 106 catchments. Diamonds denote catchments having a constant-stage tanh-type curve, the colored bar indicates the average value of a as total rainfall loss under saturation conditions (mm), triangles denote catchments having a non-constant-stage tanh-type curve, and circles denote other catchments that lack complete datasets.....	7
Figure 2.2. Percentage of data availability for some catchments in Japan.....	8
Figure 2.3. Water holding capacity distribution profile.....	12
Figure 2.4. Schematic diagram of multi-layer model	18
Figure 2.5. Catchment area of Hachisu dam catchment in Mie Prefecture.....	19
Figure 3.1. Hydrograph separation by using local minimum method for the Jyouzankei dam catchment in Hokkaido Prefecture, Japan. The direct runoff is denoted by solid red line whereas the base flow is denoted by solid blue line.....	21
Figure 3.2. Number of rainfall events having total rainfall in the range of every 50 mm for each catchment.....	22
Figure 3.3. Relationship between total rainfall (mm) and total rainfall loss (mm) for the (a) Kusaki dam catchment in Gunma Prefecture, (b) Kyuuragi dam catchment in Saga Prefecture, (c) 23 catchments in Japan having a constant-stage tanh-type curve.....	23
Figure 3.4. (a) Standard deviation values to explain about initial water amount, (b) Maximum and minimum values of runoff parameters to explain about extreme cases.....	29
Figure 3.5. Geographic distribution of standard deviation of total rainfall loss under saturation conditions for 23 catchments having a constant-stage tanh-type curve.....	29
Figure 4.1. Average frequency of rainfall events for each range of total rainfall. The blue bars denote catchments having a constant-stage tanh-type curve, and the red bars denote catchments having a non-constant-stage tanh-type curve.....	32
Figure 4.2. Comparison between simulated and observed discharges of the target rainfall for the Kusaki dam catchment in Gunma Prefecture, Japan.....	36
Figure 4.3. Relationship between average values of ΔQ_p (%) and Δt_p (h) with the corresponding standard deviation values for the Kusaki dam catchment in Gunma Prefecture, Japan, based on an ensemble of 26 rainfall events.....	37

Figure 5.1. Geographic distribution of 47 catchments. Diamonds denote catchments having a constant-stage tanh-type curve, the colored bar indicates the value of the minimum total rainfall required to generate direct runoff (mm), triangles denote catchments having a non-constant-stage tanh-type curve, and circles denote other catchments that lacked complete datasets.....	40
Figure 5.2. Geographic distribution of rock class area in km ² for (a) igneous rock class, (b) metamorphic rock class, (c) sedimentary rock class.....	56
Figure 5.3. Scatter diagram between fraction of catchment area (km ²) and fraction of the minimum total rainfall required to generate direct runoff (mm) for catchments having a constant-stage tanh-type curve. Red denotes igneous rocks, green denotes metamorphic rocks, and purple denotes sedimentary rocks.....	57
Figure 6.1. Relationship between total rainfall and total rainfall loss of 103 cases for the target rainfall.....	72
Figure 6.2. Water holding capacity distribution of 103 cases for the target rainfall.....	73
Figure 6.3. Outflow contribution rate of 103 cases for the target rainfall.....	73
Figure 6.4. Effective rainfall intensity of 103 cases for the target rainfall.....	74
Figure 6.5. Runoff simulation of 103 cases for the target rainfall.....	74
Figure 6.6. Probability density function (PDF) of peak runoff height. Blue line denotes PDF of peak runoff height based on uncertainty of rainfall distribution, and red line denotes PDF of peak runoff height based on uncertainty of rainfall distribution and uncertainty of water holding capacity. The uncertainty of rainfall distribution is considered for different cases: a) 5% b) 10% c) 15% d) 20%.....	75
Figure 6.7. The hydrograph of case 1 rainfall	77
Figure 6.8. Probability density function (PDF) of peak runoff height based on uncertainties associated with uncertainty of rainfall distribution and water holding capacity. The uncertainty associated with rainfall distribution is simulated based on 3 different cases i.e. 10% (green line), 15% (red line), and 20% (purple line).....	79
Figure 6.9. Shape of hyetograph is classified into 3 types of triangle a) triangle with peak come earlier b) isosceles triangle c) triangle with peak come later.....	80
Figure 6.10. 1 to 1 plot between uncertainty of peak runoff associated with uncertainty of water holding capacity and uncertainty of peak runoff associated with uncertainty of rainfall distribution for different uncertainty (a) 10%, (b) 15%, and (c) 20%.	82

LIST OF TABLES

Table 2.1. List of 36 catchments in 14 prefectures in Japan.....	8
Table 3.1. Summary of values of runoff parameters a and b including their standard deviation values.....	23
Table 3.2. Summary of seasonal mean precipitation (mm), and total rainfall required to cause saturation conditions (mm) for 23 catchments having a constant-stage tanh-type curve.....	27
Table 4.1. Parameter values for a as total rainfall loss under saturation conditions (mm) and b based on the inverse method for different ranges of total rainfall.....	35
Table 4.2. Results of ΔQ_p and Δt_p of the target rainfall for Kusaki dam catchment in Gunma Prefecture, Japan.....	37
Table 5.1. Correlation coefficient between catchment morphometric parameters and a) total rainfall loss under saturation condition, b) minimum total rainfall required to generate direct runoff, and c) Dirac's delta function.....	43
Table 5.2. Rock classification for some catchments in Japan.....	44
Table 5.3. Percentage area of rock classes for each catchment.....	56
Table 6.1. Summary of uncertainty of peak runoff height.....	76
Table 6.2. Summary of selected big rainfall events occurred at Kusaki dam catchment.....	76
Table 6.3. Summary of uncertainty of peak runoff height associated with uncertainties of rainfall distribution and water holding capacity.....	79
Table 6.4. Summary of uncertainty of peak runoff associated with uncertainties of rainfall distribution and water holding capacity.....	80
Table 6.5. Summary of uncertainty of peak runoff based on each uncertainties and its shape of hyetograph	81

Chapter 1. INTRODUCTION

1.1 RESEARCH BACKGROUND

Simulating runoff with higher accuracy to reduce the flood risk is one of the main research targets in hydrology. Rain water falling to the ground surface will either run off along the surface or infiltrate into the soil, which is highly influenced by the soil permeability that affects the infiltration capacity. When the rainfall rate is larger than the infiltration capacity, the excess rainfall flows over the surface causing flooding and erosion (Brutsaert, 2005). Thus, excess rainfall or effective rainfall is the portion of rainfall that contributes to direct runoff that causes flooding. Effective rainfall causes build-up of pore-water pressure, which weakens the materials supporting the slope, thereby causing landslides and flows (e.g., creep, debris flow, and debris avalanche) (Pipkin et al., 2005). Effective rainfall can be estimated by estimating the potential water storage capacity in a target catchment. By estimating effective rainfall, rainfall loss as the parameter that indicates the portion of rainfall that does not contribute to direct runoff can be calculated. Rainfall loss that is defined as the difference between the observed total rainfall and the effective rainfall, consists mainly of infiltration with some allowance for interception, evapotranspiration, subsurface flow into neighboring basins, and depression storage (Chow et al., 1988). Thus, the infiltrated rainfall as the main contributor to the rainfall loss can be used to estimate the potential catchment storage.

An experiment by considering only the effects of soil moisture content and evaporation for understanding the runoff phenomenon showed that the loss of infiltrated rainfall is closely correlated with the initial soil moisture content (Hino et al., 1988). Other studies related to effective rainfall estimation also concluded that rainfall loss associated with infiltration is highly correlated with the initial soil moisture content (Mezencev, 1948; Philip, 1957; MIs, 1980; Jakeman et al., 1990; Post and Jakeman, 1999). These studies merely make use of a data-based approach, where the losses are analyzed by empirical processing of available observations (primarily rainfall and river flow time series). Some authors even recommend or apply such approach to identify several rainfall-runoff sub-processes and construct conceptual rainfall-runoff models in a data-based, top-down based way (e.g. Klemeš, 1983; Sivapalan et al., 2003; Fenicia et al., 2007; Willems, 2014). The main disadvantage of these approaches is that extensive data is available for the catchment under study.

Previous studies have proposed estimation of potential catchment storage by using different hydrological models. Hydrological models can be classified into two groups, lumped models and distributed models. Performance comparisons between lumped and distributed models have been studied (Boyle et al., 2001; Ajami et al., 2004; Andreassian et al., 2004; Carpenter et al., 2006; Das et al., 2008). In lumped models, a catchment is regarded as a single and homogeneous unit, so the inflow is routed to the outlet using a single-unit hydrograph. Lumped models require less forcing of input data, and runoff simulation can be simplified as a function of time only. The main characteristic of distributed models is their attention to spatial variability in forcing input data and catchment morphometric parameters, and more forcing of input data is required in simulations. For application purposes, especially for discharge prediction in ungauged basins (PUB), lumped models are preferred

over distributed models due to their simplicity because they require less forcing of the input data.

A more specific study to estimate rainfall loss by using a lumped model was proposed. The model is based on total rainfall and total rainfall loss relationship (Yamada and Yamazaki, 1983). The relationship between total rainfall and total rainfall loss is well fitted using the tanh function fitting curve. By taking the second derivative of the relationship between the total rainfall and the total rainfall loss, the water holding capacity distribution can be obtained. Therefore, the theory of water holding capacity distribution is used to analyze a catchment's capacity to hold rainfall that does not contribute to direct runoff. The main benefit of this theory lies in the limited amount of input data required, i.e. observed rainfall and runoff data only. The proposed theory was further developed to estimate effective rainfall intensity, and to clarify the nonlinearity of runoff phenomena using a lumped model (Yamada, 2003).

A subsequent study by Kure and Yamada (2004) aimed at estimating effective rainfall intensity based on the water holding capacity using an inverse approach. It showed acceptable results after application to the runoff simulation at the Kusaki dam catchment in Gunma Prefecture, Japan (36.54°N, 139.37°E), by using continuous hourly rainfall and runoff for 20 years. However, the runoff was simulated by using runoff parameters which were obtained deterministically.

In this study, the hourly rainfall and hourly runoff data are obtained from the Ministry of Land, Infrastructure, Transportation, and Tourism (MLIT), Japan database during summer and autumn (June-October) at least for 10 years (2002-2011). By checking the data quality, among 106 catchments available in the database, only 36 catchments have the continuous data set that fulfill the target period in this study. Thus, total rainfall and total rainfall loss relationship is applied to those 36 catchments, and results indicate that those 36 catchments can be classified into 2 groups i.e. 23 catchments having a constant-stage tanh-type curve, and 13 catchments having a non-constant-stage tanh-type curve. Based on the physical interpretations given before to the linear and constant parts of the tanh curves, catchments having a constant-stage tanh-type curve are characterized by a constant stage after the linear stage due to some heavy rainfall events that have small total rainfall loss, which catchments having a constant-stage tanh-type curve demonstrate saturation conditions (Supraba and Yamada, 2014).

The total rainfall and total rainfall loss is related by two important runoff parameters, namely, parameter a and parameter b . From the obtained results, this study found that runoff parameter a in the total rainfall and total rainfall loss relationship represents the height of tanh curve that can be used to estimate the potential catchment storage for catchments having a constant-stage tanh-type. Thus, parameter a is an important parameter that can be used to estimate the capacity of a catchment to hold or to store water during a rainfall event. By knowing the capacity of a catchment, the amount of rain water that become direct runoff that causes flooding can be estimated (Supraba and Yamada, 2014).

However, the plotting result of total rainfall and total rainfall loss relationship show that similar values of total rainfall occurred in two different rainfall events have different values of total rainfall loss. This difference indicates the effect of initial soil moisture condition. The rainfall event that having

bigger total rainfall loss means the respective rainfall occurred when the catchment was initially dry, and on contrary, the rainfall event that having smaller total rainfall loss means the respective rainfall occurred when the catchment was initially wet. The plotting result of total rainfall-total rainfall loss relationship also show the standard deviation values (1σ) of runoff parameters a and b . Thus, this study interpret the value of 1σ of runoff parameter a as the initial water amount that reflects the initial soil moisture condition, and is used as the parameter to explain about the uncertainty associated with water holding capacity (initial water amount in a catchment) (Supraba and Yamada, 2014; 2015).

The limitation of the water holding capacity distribution theory is that the runoff parameters in the relationship between total rainfall and total rainfall loss only can be used to estimate effective rainfall intensity for a target catchment that has ever experienced saturation conditions. The saturation conditions in the region are mainly due to heavy rainfall events with small total rainfall loss. Thus, this study proposes two different methods, namely, localized gradient method and inverse method to estimate runoff parameters for catchments that have never experienced the saturation condition (catchments having a non-constant stage tanh-type curve).

The mentioned theory is based on an assumption that rainfall as input data is true, when actually rainfall contains uncertainty. In Japan, in-situ observation such as the Automated Meteorological Data Acquisition System (AMeDAS) Japan Meteorological Agency (JMA) and Ministry of Land, Infrastructure, Transportation, and Tourism (MLIT) rain gauges cover the whole Japan. Rain gauges measure the rainfall intensity near to the land surface, but there is limitation on their spatial representativeness due to the location and density of rain gauges. The measured amounts are influenced by several factors such as wind, snowfalls, station relocation, and change of the sensors (Burcea et al., 2012). Rain gauges based rainfall intensity measurements can be biased by factors like wind and evaporation in the range of 10-20% (Cheval et al., 2011).

Japan Meteorological Agency (JMA) estimates rainfall intensity by using C-band radar and X-band radar that having high spatial and temporal resolution over extended areas. However, radar also has uncertainty due to several factors such as hardware calibration, mountain blockage, and anomalous propagation (Yilmaz et al., 2005). Villarini et al. (2008) stated that remote sensing contains uncertainties due to lacking of knowledge to fully understand the physical processes, parameter estimation, and the device measurement. Thus, whether measured directly by rain gauges or indirectly by remote sensing techniques, all rainfall intensity measurement contain uncertainty (Yilmaz et al., 2005; Villarini et al., 2008).

The theory about uncertainty is analogous to the random term in Brownian motion. The first theory of Brownian motion is in consequence of the role of Gaussian variables in probability (Einstein, 1956). The stochastic force and derivation of Einstein's theory of Brownian motion from Newton's second law was introduced by Paul Langevin (Li and Raizen, 2013). Yoshimi et al. (2015) investigated the uncertainty of peak runoff height using the relationship between Ito stochastic differential equation (sde) and Fokker-Planck equation, and concluded that 10% uncertainty of rainfall distribution contributes to the uncertainty of peak runoff height.

1.2 RESEARCH OBJECTIVE

The main objective of this study is to investigate the uncertainty of runoff by considering two independent uncertainties i.e. uncertainty associated with water holding capacity, and uncertainty associated with rainfall distribution. Two different methods, ensemble method and stochastic differential equation method, are proposed to quantify the uncertainty of runoff associated with those two uncertainties.

Chapter 2. METHODOLOGY

2.1 DATA

Hourly rainfall and hourly runoff data of 106 catchments in Japan are obtained from the Water Information System database of the Ministry of Land, Infrastructure, Transportation, and Tourism of Japan. Runoff refers to the dam inflow (inlet flow). The geographical distribution of all catchments is shown in **Figure 2.1**. The meaning of each symbol will be explained in the following chapter.

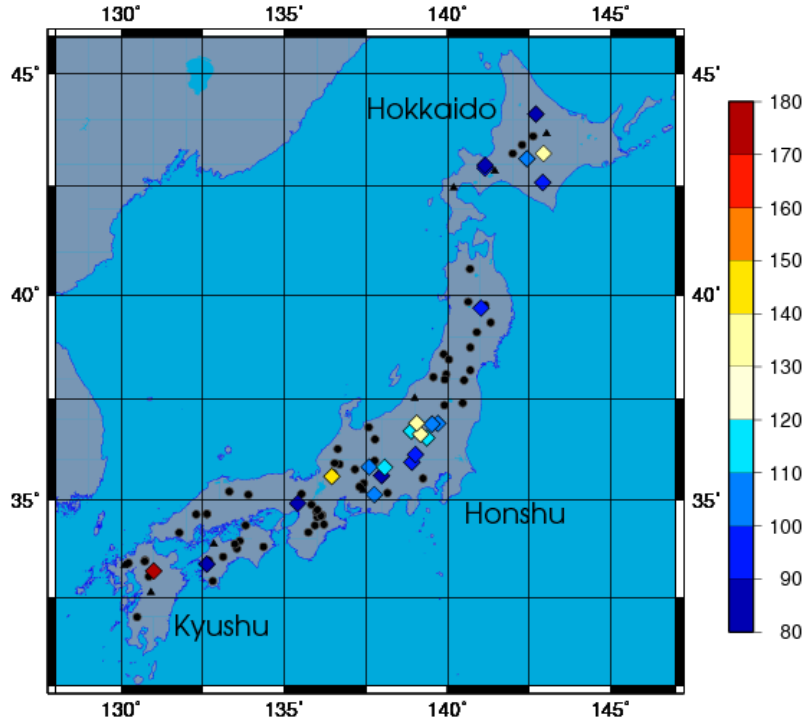


Figure 2.1. Geographic distribution of 106 catchments. Diamonds denote catchments having a constant-stage tanh-type curve, the colored bar indicates the average value of a as total rainfall loss under saturation conditions (mm), triangles denote catchments having a non-constant-stage tanh-type curve, and circles denote other catchments that lack complete datasets.

The target period of this study is the summer and autumn seasons, i.e., June to October, at least for 10 years (2002–2011). Among the 106 catchments are some for which data of the full target period are lacking; some have data only since June 2005 or June 2008, some have no rainfall data but have runoff data, and some have unrealistic rainfall values. Such catchments are excluded from the analysis. **Figure 2.2** shows the percentage of data availability for some catchments. Additionally, the shapes of the hydrographs vary among catchments due to snow melt, and snow-melt periods are also excluded from the analysis.

After carefully checking the data quality, only 36 catchments located in 14 prefectures in Japan have the continuous data set. Among these 36 catchments, longer data sets could be obtained for 10 catchments located in Hokkaido Prefecture, which is the second largest and northernmost island in Japan (14 years, 1998–2011), and for the Kusaki dam catchment in Gunma Prefecture (30 years, 1982–2011) (see **Figure 2.1**). The list of 36 catchments is presented in **Table 2.1**.

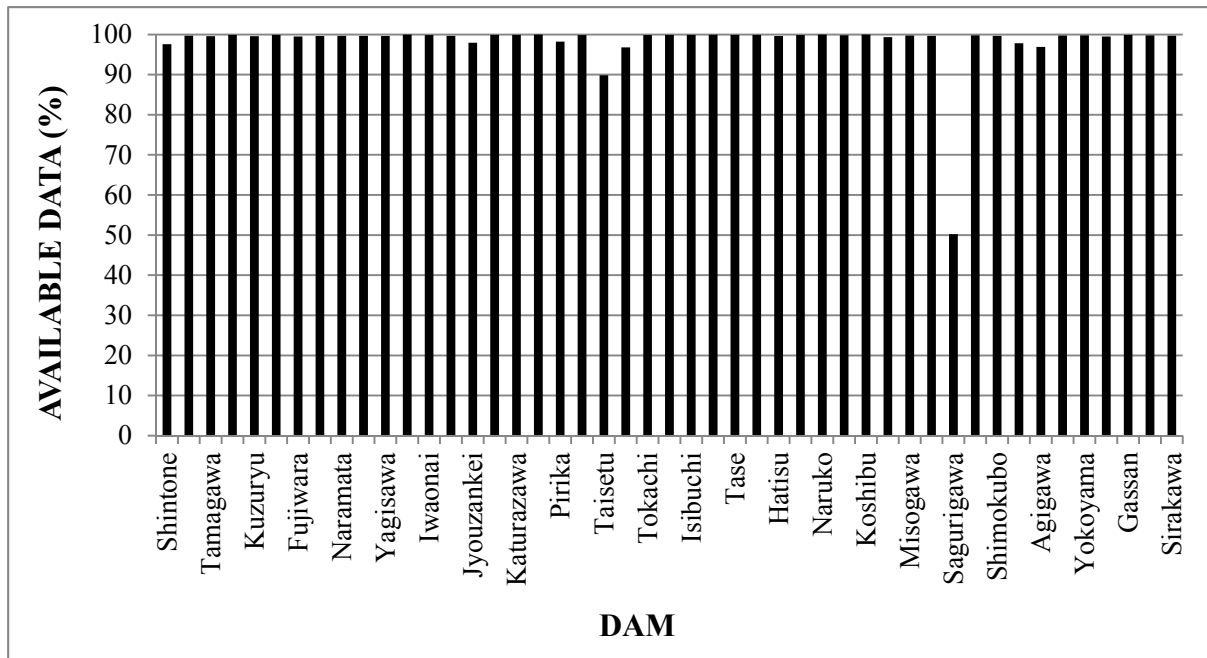


Figure 2.2. Percentage of data availability for some catchments in Japan.

Table 2.1. List of 36 catchments in 14 prefectures in Japan.

No.	Dam	Prefecture (Latitude, Longitude)	Catchment Area (km ²)
1	Houheikyuu	Hokkaido (42.92, 141.15)	136.1
2	Iwaonai	Hokkaido (44.12, 142.71)	341.6
3	Izarigawa	Hokkaido (42.85, 141.45)	113.3
4	Jyouzankei	Hokkaido (42.98, 141.16)	103.6
5	Kanayama	Hokkaido (43.13, 142.44)	410.8
6	Nibutani	Hokkaido (42.63, 142.15)	1155.5
7	Pirika	Hokkaido (42.47, 140.19)	114.4
8	Satunaigawa	Hokkaido (42.59, 142.92)	116.6
9	Taisetsu	Hokkaido (43.68, 143.04)	289.3
10	Tokachi	Hokkaido (43.24, 142.94)	598.2
11	Gosho	Iwate (39.69, 141.03)	635.2
12	Sagurigawa	Niigata (37.52, 139.00)	61.4
13	Ikari	Tochigi (36.90, 139.71)	271.2
14	Kawaji	Tochigi (36.90, 139.69)	320.7
15	Kawamata	Tochigi (36.88, 139.52)	179.4
16	Aimata	Gunma (36.71, 138.89)	110.8
17	Fujiwara	Gunma (36.8, 139.04)	400.2

No.	Dam	Prefecture (Latitude, Longitude)	Catchment Area (km ²)
18	Kusaki	Gunma (36.54, 139.37)	263.9
19	Naramata	Gunma (36.88, 139.08)	95.4
20	Simokubo	Gunma (36.13, 139.02)	323.7
21	Sonohara	Gunma (36.64, 139.18)	601.1
22	Yagisawa	Gunma (36.91, 139.06)	165.5
23	Futase	Saitama (35.94, 138.91)	170.6
24	Koshibu	Nagano (35.61, 137.98)	289.6
25	Makio	Nagano (35.82, 137.60)	307.8
26	Miwa	Nagano (35.81, 138.08)	311.0
27	Maruyama	Gifu (35.47, 137.17)	2409.0
28	Yokoyama	Gifu (35.59, 136.46)	470.7
29	Sintoyone	Aichi (35.13, 137.76)	111.4
30	Yahagi	Aichi (35.24, 137.42)	504.6
31	Hitokura	Hyogo (34.91, 135.41)	115.1
32	Ishitegawa	Ehime (33.88, 132.84)	72.6
33	Nomura	Ehime (33.36, 132.63)	168.0
34	Kyuuragi	Saga (33.33, 130.10)	33.7
35	Matsubara	Ooita (33.19, 130.99)	491.0
36	Midorikawa	Kumamoto (32.63, 130.91)	359.0

2.2 STREAMFLOW HYDROGRAPH SEPARATION

Two-component hydrograph separation methods separating streamflow hydrographs into fast and slow components, often related to surface and groundwater runoff, have been proposed since long (Linsley and Köhler, 1958). In the late 1960, methods were introduced for separation of flow hydrographs into time source components of event and pre-event water using stable isotope tracers. This type of separation, which is based on components of the water itself, introduced as a paradigm shift in how hydrologists conceptualized runoff generation (Klaus and McDonnell, 2013). Later on, other hydrograph separation techniques using recursive digital filters to separate between the rapidly occurring discharge components such as direct runoff, and the slowly changing discharge such as interflow and groundwater, were introduced. They depend on measured stream hydro-geochemistry such as natural tracers (e.g. Cl^- or SO_4^{2-}); sediment concentrations (e.g. Total Suspended Solid (TSS), turbidity) or nutrients (e.g. NO_3^-) (Rimmer and Hartmann, 2014).

Because this study aimed at testing a method that can be easily duplicated for other upper catchments, a simple model that can also work if limited observation data are available is selected.

Several such methods have been developed since the 1950s (e.g. Nathan and McMahon, 1990; Chapman, 1999; Arnold & Allen, 1999; Eckhardt, 2008; Willems, 2009). One such method, selected for this study, is the method to separate streamflow hydrographs into base flow and direct runoff components introduced by the U.S. Geological Survey (USGS). The technique is called the local minimum method (Pettyjohn and Henning, 1979; Sloto and Crouse, 1996). It starts from daily river flow time series data. The duration of direct runoff is estimated from the empirical relation:

$$N = A^{0.2} \quad (1)$$

where N is the number of days after direct runoff ceases, and A is the catchment area (mi^2). The interval used for hydrograph separation is denoted by $2N_*$, where $2N_*$ is the odd integer between 3 and 11 that is nearest to $2N$. The basis of the local minimum method is to separate the hydrographs based on the lowest discharge within moving intervals of length:

$$I = 0.5(2N_* - 1) \quad (2)$$

Based on the identified local minima, adjacent local minima are connected by straight lines to define the base flow values for each day between two local minima. This means that the base flow values are identified by linear interpolation. The hydrograph periods then start $2N_*$ days before the starting date of the separation and end $2N_*$ days after the ending date of the separation, to make sure that the rainfall event that causes the direct runoff hydrograph takes part of the interval (Pettyjohn and Henning, 1979; Sloto and Crouse, 1996).

2.3 LUMPED MODEL

(Yamada and Yamazaki, 1983; Yamada, 2003) proposed the basic equations to estimate effective rainfall intensity based on total rainfall and total rainfall loss relationship.

Total rainfall loss, (R) , can be expressed by the linear Volterra-type integral equation of the first kind:

$$F(R) = R - \int_0^R (R - h)S(h)dh \quad (3)$$

where $F(R)$ is total rainfall loss (mm), R is total rainfall (mm), and $S(h)$ is the water holding capacity distribution (1/mm). The solution can be obtained by Laplace transformation. The Laplace Transform is to transform $f(t)$ to a function $F(s)$ with complex argument s :

$$F(s) = \int_0^\infty f(t)e^{-st}dt \quad (4)$$

Solving Eq. (3) by using Eq. (4), gives the following equation:

$$S(R) = -\frac{d^2 F}{dR^2} + \left[1 - \frac{dF}{dR}\right]_{R=0} \delta(R) \quad (5)$$

A previous study suggested that for mountainous catchments the relationship between total rainfall and total rainfall loss can be well fitted using the tanh function (Kure and Yamada, 2004):

$$F(R) = a \tanh(bR) \quad (6)$$

where $\delta(R)$ is the Dirac's delta function, and a and b are runoff parameters.

From Eq. (6), a and b are the important parameters that influence the value of $F(R)$. $F(R)$ as total rainfall loss mainly express the portion of rainfall that infiltrates into the ground. The higher infiltration rate reflects the higher catchment permeability. Thus, the bigger values of a and b indicate the higher catchment permeability, and ab represents the fraction of permeable area in the catchment.

After substituting Eq. (6) into Eq. (5):

$$S(R) = -\frac{d^2(a \tanh(bR))}{dR^2} + \left[1 - \frac{d(a \tanh(bR))}{dR}\right]_{R=0} \delta(R) \quad (7)$$

And solving the derivative equation of Eq. (7):

$$S(R) = (2ab^2 \operatorname{sech}[bR]^2 \tanh[bR]) + (1 - ab)\delta(R) \quad (8)$$

The following “water-holding capacity distribution” equation is obtained after further rearrangement:

$$S(R) = (1 - ab)\delta(R) + 2ab^2 \frac{\sinh(bR)}{\cosh^3(bR)} \quad (9)$$

$\delta(R)$ determines the existence of $1 - ab$ in the 1st term of right hand side of Eq. (9) which can be explained as follows:

- When there is no rainfall, $R = 0$, thus $\delta(R) = 1$. Thus, the water holding capacity distribution is expressed as:

$$S(R) = (1 - ab)\delta(R) + 2ab^2 \frac{\sinh(bR)}{\cosh^3(bR)}$$

- When there is rainfall, $R > 0$, thus $\delta(R) = 0$. Thus, the water holding capacity distribution is expressed as:

$$S(R) = 2ab^2 \frac{\sinh(bR)}{\cosh^3(bR)}$$

Therefore, $\delta(R)$ represents the term of $1 - ab$. Thus, $\delta(R)$ is interpreted as the fraction of impermeable area in the catchment.

By integrating Eq. (9), the outflow contribution rate $ocr(t)$ can be obtained, which is defined as the rate portion of rainfall that contributes to direct runoff:

$$ocr(t) = 1 - ab \operatorname{sech}[b \operatorname{acr}(t)] \quad (10)$$

where $\operatorname{acr}(t)$ is the accumulation of hourly rainfall intensity in every time step until it reaches total rainfall R . If the outflow contribution rate is equal to 1, it means that the initially catchment condition is saturated, hence that there is no infiltration and all rainfall becomes direct runoff. Finally, the effective rainfall intensity can be calculated as follows:

$$r_e(t) = ocr(t) r(t) \quad (11)$$

where $r_e(t)$ is the effective rainfall intensity (mm/h), $r(t)$ is hourly rainfall intensity (mm/h).

The example of water holding capacity distribution profile by applying Eq. (9) is shown in **Figure 2.3**.

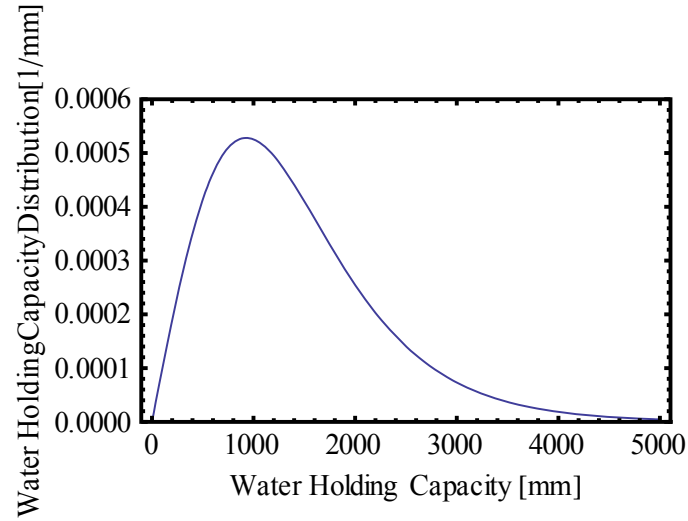


Figure 2.3. Water holding capacity distribution profile.

After obtaining effective rainfall intensity, then it is used as the input data to simulate runoff height. The basic equations to simulate runoff height can be derived as follows:

$$v = \alpha h^m \quad (12)$$

$$q = vh \quad (13)$$

where q is discharge per unit width (mm^2/h), v is surface flow velocity (mm/h), h is water ponding

depth in the surface soil layer (mm), $\alpha = \frac{k_s i}{dD^m w^{m+1}}$, where k_s is saturated hydraulic conductivity (mm/h), i is slope gradient, dD is soil thickness (mm), w is effective porosity, and m is a resistance coefficient.

Combining Eq. (12) and Eq. (13), the following equation relating discharge per unit width to water ponding depth is obtained:

$$q = \alpha h^{m+1} \quad (14)$$

The mass conservation equation equals:

$$\frac{\partial h}{\partial t} + \frac{\partial q}{\partial x} = r_e(t) \quad (15)$$

where r_e is the effective rainfall intensity (mm/h), t is time (h), and x is the unit length of the water path along the bottom slope (m).

From Eq. (14):

$$h^{m+1} = \frac{q}{\alpha} \quad (16)$$

From Eq. (16):

$$h = \left(\frac{q}{\alpha}\right)^{\frac{1}{m+1}} \quad (17)$$

After substituting Eq. (17) into Eq. (15):

$$\frac{\partial \left(\frac{q}{\alpha}\right)^{\frac{1}{m+1}}}{\partial t} + \frac{\partial q}{\partial x} = r_e(t) \quad (18)$$

From Eq. (18):

$$\frac{1}{\alpha^{\frac{1}{m+1}}} \frac{\partial q^{\frac{1}{m+1}}}{\partial t} + \frac{\partial q}{\partial x} = r_e(t) \quad (19)$$

After multiplying Eq. (19) by $q^{\frac{m}{m+1}}$:

$$q^{\frac{m}{m+1}} \frac{1}{\alpha^{\frac{1}{m+1}}} \frac{\partial q^{\frac{1}{m+1}}}{\partial t} + q^{\frac{m}{m+1}} \frac{\partial q}{\partial x} = q^{\frac{m}{m+1}} r_e(t) \quad (20)$$

From Eq. (20):

$$\frac{1}{\alpha^{\frac{1}{m+1}}} \frac{\partial q}{\partial t} + q^{\frac{m}{m+1}} \frac{\partial q}{\partial x} = q^{\frac{m}{m+1}} r_e(t) \quad (21)$$

After multiplying Eq. (21) by $\alpha^{\frac{1}{m+1}}$:

$$\frac{\partial q}{\partial t} + \alpha^{\frac{1}{m+1}} q^{\frac{m}{m+1}} \frac{\partial q}{\partial x} = \alpha^{\frac{1}{m+1}} q^{\frac{m}{m+1}} r_e(t) \quad (22)$$

Eq. (22) is the Kinematic Wave equation.

Rearranging Eq. (22):

$$\frac{\partial q}{\partial t} + a q^{\frac{m}{m+1}} \frac{\partial q}{\partial x} = a q^{\frac{m}{m+1}} r_e(t) \quad (23)$$

From Eq. (22) and Eq. (23):

$$a = (m + 1) \alpha^{\frac{1}{m+1}} \quad (24)$$

In this lumped model, the slope length is not the real slope length of the catchment, because only discharge in the vicinity of the river (nearby tributaries) contributes to the direct runoff. Hence, the discharge can be considered as a function of time only. For this reason, the partial differential equation $q(x, t)$ can be transformed into the ordinary differential equation as follows:

$$q(x, t) \approx x q_*(t) \quad (25)$$

After substituting Eq. (25) into Eq. (23):

$$\frac{dxq_*(t)}{dt} + a_0 x q_*(t)^{\frac{m}{m+1}} \frac{dxq_*(t)}{dx} = a_0 x q_*(t)^{\frac{m}{m+1}} r_e(t) \quad (26)$$

After rearranging Eq. (26):

$$\frac{xdq_*(t)}{dt} + a_0 x q_*(t)^{\frac{m}{m+1}} q_*(t) = a_0 x q_*(t)^{\frac{m}{m+1}} r_e(t) \quad (27)$$

After rearranging Eq. (27):

$$\frac{xdq_*(t)}{dt} = a_0 x q_*(t)^{\frac{m}{m+1}} r_e(t) - a_0 x q_*(t)^{\frac{m}{m+1}} q_*(t) \quad (28)$$

After rearranging Eq. (28):

$$\frac{xdq_*(t)}{dt} = a_0 x q_*(t)^{\frac{m}{m+1}} (r_e(t) - q_*(t)) \quad (29)$$

After dividing Eq. (29) by x :

$$\frac{dq_*(t)}{dt} = a_0 q_*(t)^{\frac{m}{m+1}} (r_e(t) - q_*(t)) \quad (30)$$

After rearranging Eq. (30):

$$\frac{dq_*(t)}{dt} = a_0 q_*(t)^\beta (r_e(t) - q_*(t)) \quad (31)$$

where q_* is the runoff rate or runoff height (mm/h), L is slope length (mm), and β is one of the runoff parameters which is a function of resistance coefficient, m . Eq. (31) is the well-known runoff height equation showing runoff flow on a slope which is widely used for researches and practical

purposes in the mountain catchments where $a_0 = aL^{\beta-1} = (m+1)\alpha^{\frac{1}{m+1}}L^{\frac{-1}{m+1}}$.

From Eq. (30) and Eq. (31):

$$\beta = \frac{m}{m+1} \quad (32)$$

Runoff parameters a_0 and β can be obtained from the recession curve of a stream flow hydrograph when the rain has stopped so there is no rainfall. Thus, the runoff equation during recession period can be obtained by assuming that $r_e(t)$ in Eq. (31) is equal to zero. Hence, Eq. (31) can be modified as follows:

$$\frac{dq_*(t)}{dt} = a_0 q_*(t)^\beta (-q_*(t)) \quad (33)$$

Rearranging Eq. (33):

$$\frac{dq_*(t)}{dt} = -a_0 q_*(t)^{\beta+1} \quad (34)$$

Eq. (34) is the runoff height equation during the recession period.

When β is equal to 0, the runoff is linear and Eq. (34) can be expressed as follows:

$$\frac{dq_*(t)}{dt} = -a_0 q_*(t) \quad (35)$$

By solving Eq. (35), the groundwater outflow can be expressed as follows (Werner and Sundquist, 1951; Takagi, 1966; Roche, 1974):

$$q_{*0}(t) = q_{*0} \exp(-a_0 t) \quad (36)$$

Eq. (36) is analogous to the exponential equation of a recession curve of a hydrograph that explaining the diminishing discharge from storage in the absence of further replenishment (Maillet, 1905):

$$q(t) = q_0 \exp(-a t) \quad (37)$$

where $q(t)$ is the discharge at time t , q_0 is the initial discharge, and a is constant in which usually it can be replaced by k that is the recession constant. Eq. (37) is the most widely used equation to express the part or all of the entire recession. Eq. (31) is the runoff equation for a single layer model. (Yoshimi and Yamada, 2013) proposed the multi-layer model of runoff equation by considering the vertical infiltration, sub-surface flow, and sub-merged depth based on the following concept:

- Vertical infiltration ($q_v(t)$)
If effective rainfall intensity ($r_e(t)$) < saturated hydraulic conductivity (K), then the rate of vertical infiltration ($q_v(t)$) is equal to the effective rainfall intensity ($r_e(t)$).

- Sub-surface flow ($q_s(t)$)

If sub-surface flow ($q_s(t)$) < the potential maximum value of q_s (q_c), then the sub-surface flow ($q_s(t)$) is equal to ($q_s(t)$). It means the effective rainfall that infiltrates to the ground will be sub-surface flow. Thus, if ($q_s(t)$) is equal to q_c , then the effective rainfall becomes surface runoff ($q(t)$). The equation is expressed as follows:

$$q'_s(t) = a_0 q_s(t)^\beta [q_v(t) - q_s(t)] \quad (38)$$

where q_c is calculated as follows:

$$q_c = \frac{i/n \, dD^{m+1}}{x} \quad (39)$$

and

$$n = \frac{dD^m w^{m+1}}{k_s} \quad (40)$$

where i is slope gradient, n is Manning's roughness coefficient, m is resistance coefficient, x is slope length (mm), dD is surface soil thickness (mm), and w is effective porosity.

- Surface flow ($q(t)$)

- If effective rainfall ($r_e(t)$) $\geq$ saturated hydraulic conductivity (K), and sub-surface flow ($q_s(t)$) $\geq$ the potential maximum value of q_s (q_c), then effective rainfall ($r_e(t)$) becomes surface flow ($q(t)$). The equation is expressed as follows:

$$q'(t) = a_{s0} q(t)^{\beta_s} [r_e(t) - q_s(t) - q(t)] \quad (41)$$

where a_{s0} and β_s are runoff parameters for surface runoff. The equations of a_{s0} and β_s are as follows:

$$a_{s0} = (m_s + 1) \left(\frac{\sqrt{i}}{n_s x} \right)^{\frac{1}{m_s+1}} \quad (42)$$

where m_s is the resistance coefficient for surface runoff i.e. 2/3, n_s is Manning's roughness coefficient for surface runoff. In this study, n_s is 0.03 for natural straight channel.

- If effective rainfall ($r_e(t)$) $\geq$ saturated hydraulic conductivity (K), then effective rainfall ($r_e(t)$) becomes surface flow ($q(t)$). The equation is expressed as follows:

$$q'(t) = a_{s0} q(t)^{\beta_s} [r_e(t) - q_v(t) - q(t)] \quad (43)$$

- If sub-surface flow ($q_s(t)$) $\geq$ the potential maximum value of q_s (q_c), then the sub-surface flow ($q_s(t)$) becomes surface flow ($q(t)$). The equation is expressed as follows:

$$q'(t) = a_{s0} q(t)^{\beta_s} [r_e(t) - q_s(t) - q(t)] \quad (44)$$

- Sub-merged depth ($h_s(t)$)

- If effective rainfall ($r_e(t)$) $\geq$ saturated hydraulic conductivity (K), and sub-surface flow $q_s(t) \geq$ the potential maximum value of q_s (q_c), then sub-merged depth can be expressed as follows:

$$h'_s(t) = r_e(t) - q_s(t) - q(t) \quad (45)$$

- If effective rainfall ($r_e(t)$) $\geq$ saturated hydraulic conductivity (K), then sub-merged depth can be expressed as follows:

$$h'_s(t) = r_e(t) - q_v(t) - q(t) \quad (46)$$

- If sub-surface flow ($q_s(t)$) $\geq$ the potential maximum value of q_s (q_c), then sub-merged depth can be expressed as follows:

$$h'_s(t) = r_e(t) - q_s(t) - q(t) \quad (47)$$

The diagram to explain about the parameters in the multi-layer model is shown in **Figure 2.4**.

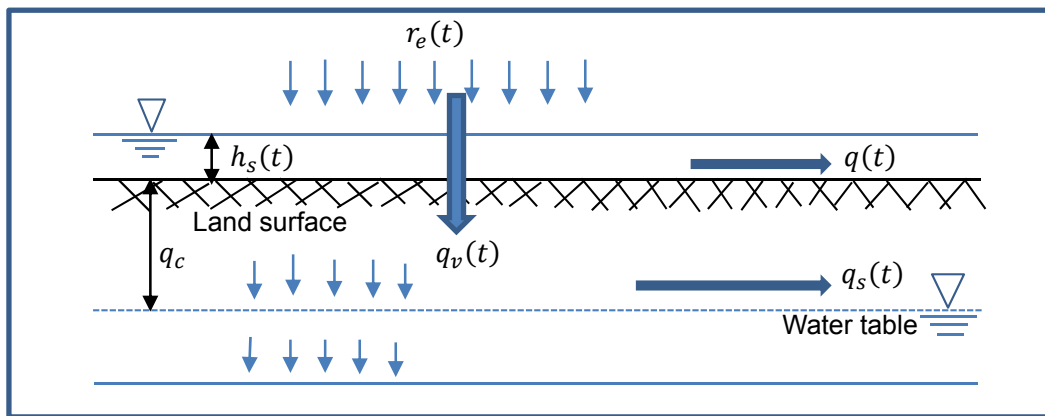


Figure 2.4. Schematic diagram of multi-layer model.

From Eq. (11), effective rainfall is a function of outflow contribution rate that is related to the outflow contribution area. The illustration of the outflow contribution area can be seen in **Figure 2.5**. The outflow contribution area is the catchment area in the vicinity of the river where only rain that falls in that area will contribute to the direct runoff in the target outlet. The other portion of rain that falls in the outside of the outflow contribution area does not contribute to direct runoff because the travelling time to reach the nearby tributaries/ streams is longer, so along the way the rain may infiltrates into the ground or evaporated or trapped in the depression storage so it does not contribute to the direct runoff.

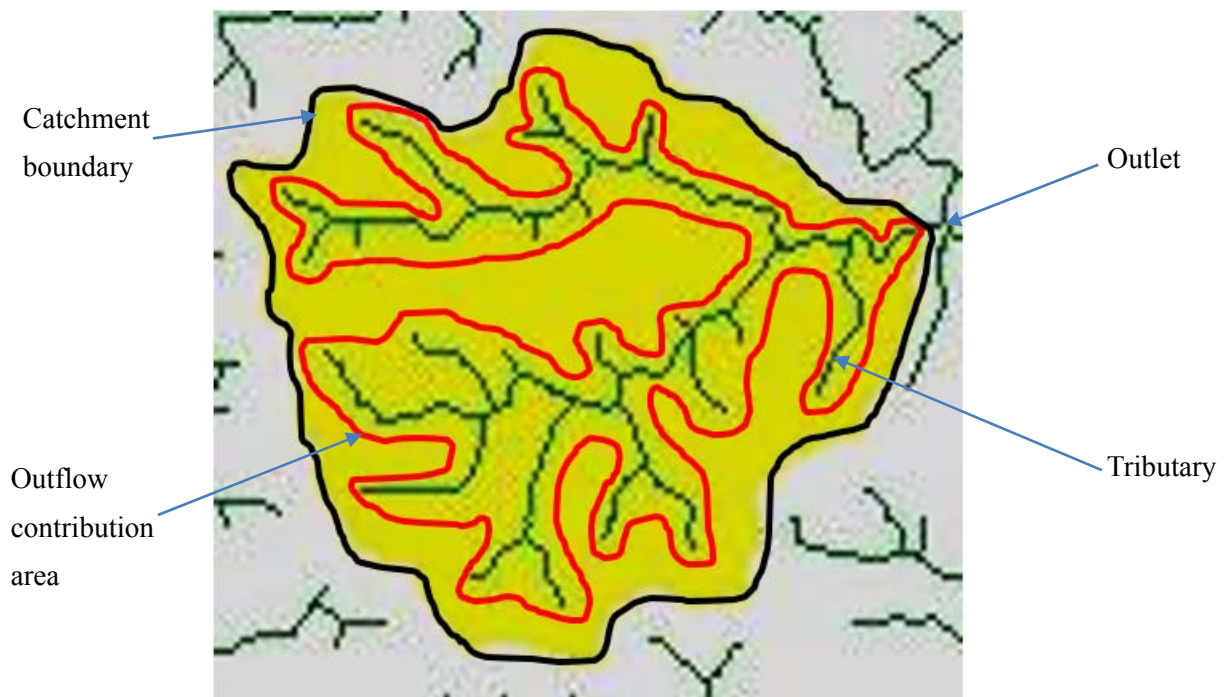


Figure 2.5. Catchment area of Hachisu dam catchment in Mie Prefecture.

Chapter 3. POTENTIAL CATCHMENT STORAGE ESTIMATION

This chapter is written based on a published paper:

Intan Supraba and Tomohito J. Yamada, “Catchment storage estimation based on total rainfall-total loss rainfall relationship for 47 catchments in Japan”, *Journal of Japan Society of Civil Engineers, Ser.B1 (Hydraulic Engineering)*, Vol. 70, No. 4, I_169-I_174, 2014.

The content of this chapter has been presented in the following conferences:

- Intan Supraba and Tomohito J. Yamada, “Catchment Storage Estimation Based on Total Rainfall-Total Loss Rainfall Relationship for 47 Catchments in Japan”, in The 58th Annual Conference on Hydraulic Engineering of Japan Society of Civil Engineers, 10th-12th March 2014, Kobe University, Japan.
- Intan Supraba and Tomohito J. Yamada, “Surface Runoff Estimation Based on Total Rainfall-Total Loss Rainfall Relationship for Catchments in Ishikari River”, in Japan Geoscience Union (JpGU) Meeting 2014, 28th April-2nd May 2014, Pacifico Yokohama, Japan.
- Intan Supraba and Tomohito J. Yamada, “Catchment Storage Estimation Based on Total Rainfall-Total Loss Relationship for 65 Catchments in Japan”, in Asia Oceania Geosciences Society (AOGS) 2014, 28th July-1st August 2014, Sapporo, Japan.

3.1 RELATIONSHIP BETWEEN TOTAL RAINFALL AND TOTAL RAINFALL LOSS

By using Eq. (2), base flow and direct runoff of each catchment are separated from the daily river flow time series for 36 catchments. After this runoff separation and the corresponding separation of the time series in streamflow hydrographs, rainfall events could be determined for direct runoff hydrograph periods. One rainfall event is defined per hydrograph period, as a period with rainfall but without rain on the hours before and after the event. The dry periods in between these rain events obviously correspond with the base flow recession periods. The number of rainfall and direct runoff events of each catchment varies because the target period, the catchment area, and runoff data of each catchment varies. For each rainfall or hydrograph event, the duration in days, and the total rainfall depth could be computed. **Figure 3.1** shows an example of the direct runoff – base flow separation for the Jyouzankei dam catchment in Hokkaido Prefecture, Japan (42.98°N, 141.16°E).

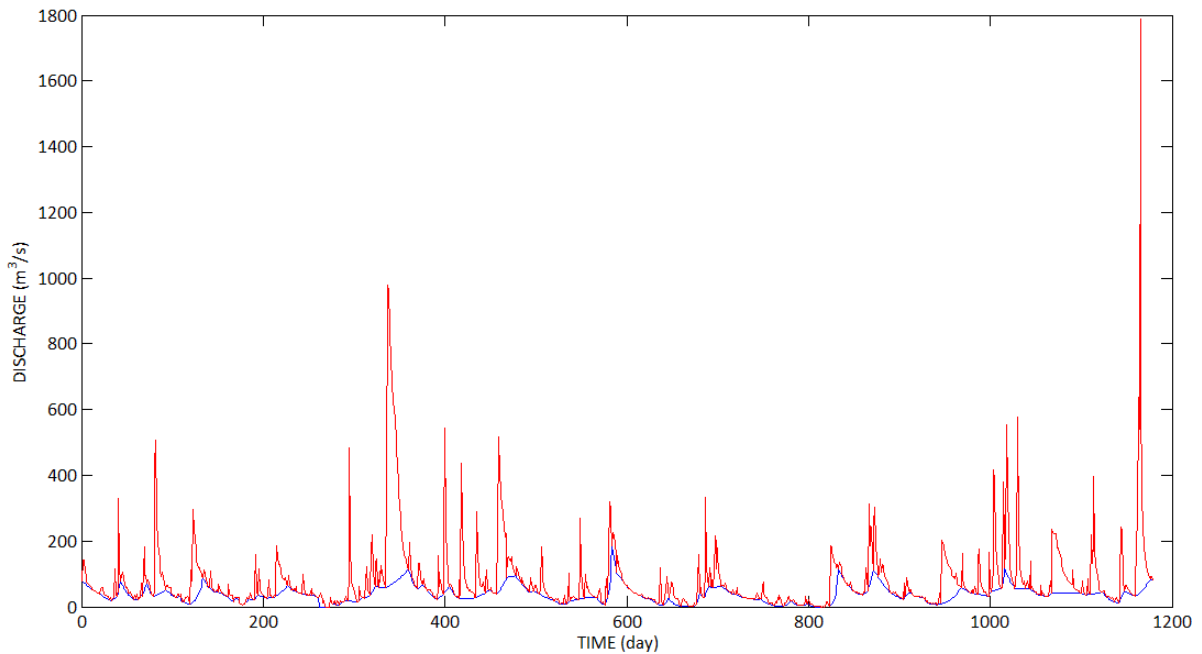


Figure 3.1. Hydrograph separation by using local minimum method for the Jyouzankei dam catchment in Hokkaido Prefecture, Japan. The direct runoff is denoted by solid red line whereas the base flow is denoted by solid blue line.

For each of the rainfall or direct hydrograph events, next to the total rainfall depth the total rainfall loss was calculated by subtracting from the total rainfall depth the total amount of direct runoff. This total rainfall loss is defined as the total amount of rainfall that does not contribute to direct runoff because of several factors such as infiltration, evaporation, interception by vegetation or surface depressions.

After calculating total rainfall and total rainfall loss for each catchment, then the summary of the number of rainfall events having total rainfall every 50 mm is shown in **Figure 3.2**.

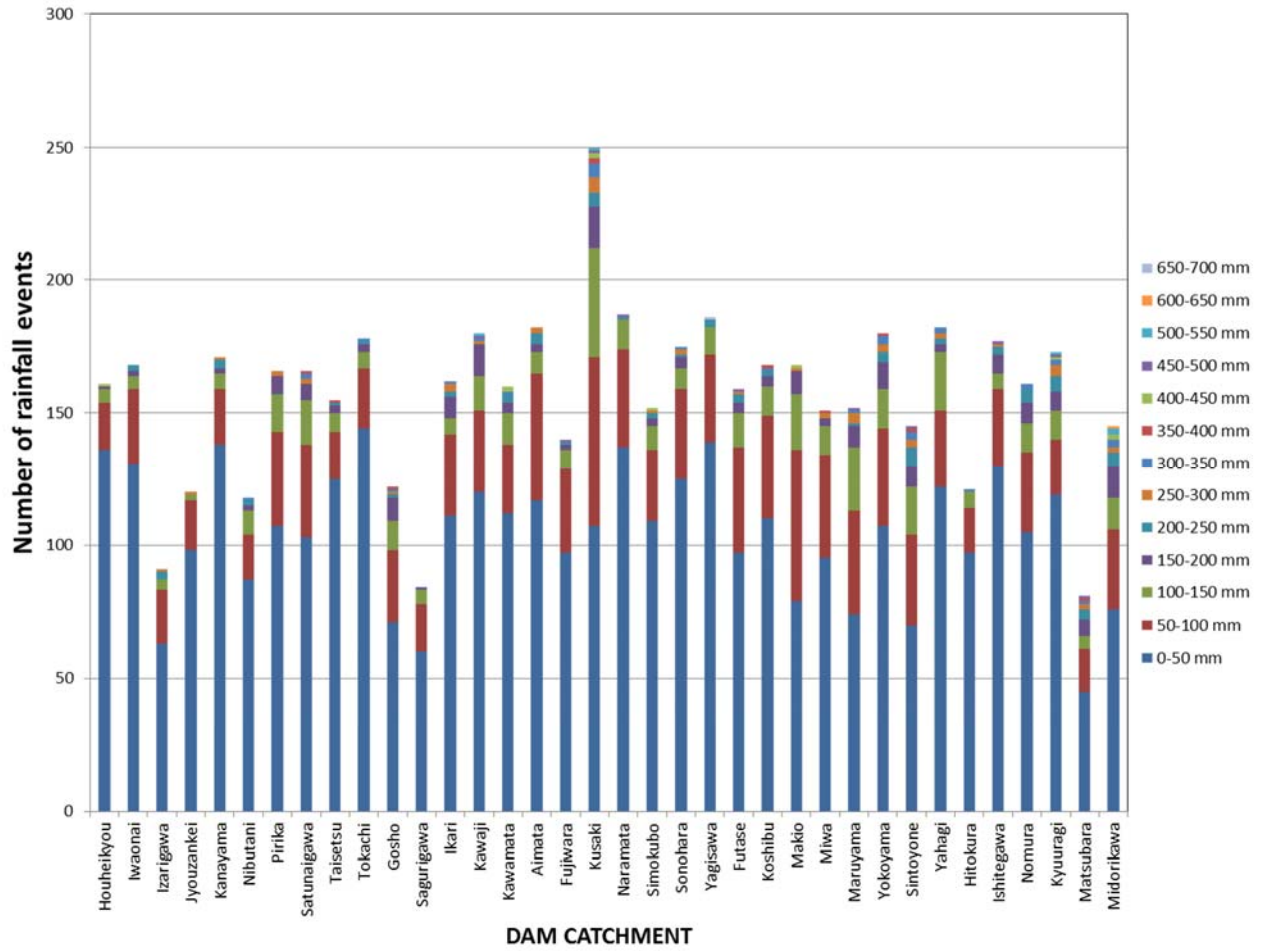


Figure 3.2. Number of rainfall events having total rainfall in the range of every 50 mm for each catchment.

Figure 3.2 shows that majority of rainfall events having total rainfall in the range of 0-50 mm, followed by total rainfall in the range of 50-100 mm. It means that majority of rainfall events occurred with small total rainfall.

After obtaining total rainfall and total rainfall loss as the output of hydrograph separation, the values of a and b can be obtained by plotting the tanh function using **Eq. (6)**. These a and b values as runoff parameters can be employed for calculating effective rainfall intensity to simulate runoff only when the plotted total rainfall–total rainfall loss relationship shows a constant stage tanh-type curve.

Thus, after applying **Eq. (6)** to obtain the relationship between total rainfall and total rainfall loss of 36 catchments, results show that those catchments can be classified into 2 groups i.e. catchments having a constant-stage tanh-type curve (23 catchments), and catchments having a non-constant-stage tanh-type curve (13 catchments). The examples of a catchment having a constant-stage tanh-type curve, a catchment having a non-constant-stage tanh-type curve, and the overplotted of 23 catchments having a constant-stage tanh-type curve are shown in **Figures 3.3. (a), (b), and (c)**, respectively. **Figure 2.1** shows that the catchments having a constant-stage tanh-type curve, denoted by diamonds, are distributed from Hokkaido to Kyushu Island. Catchments having a non-constant-stage tanh-type curve are denoted by triangles, and those catchments that lacked the complete data set are denoted by circles.

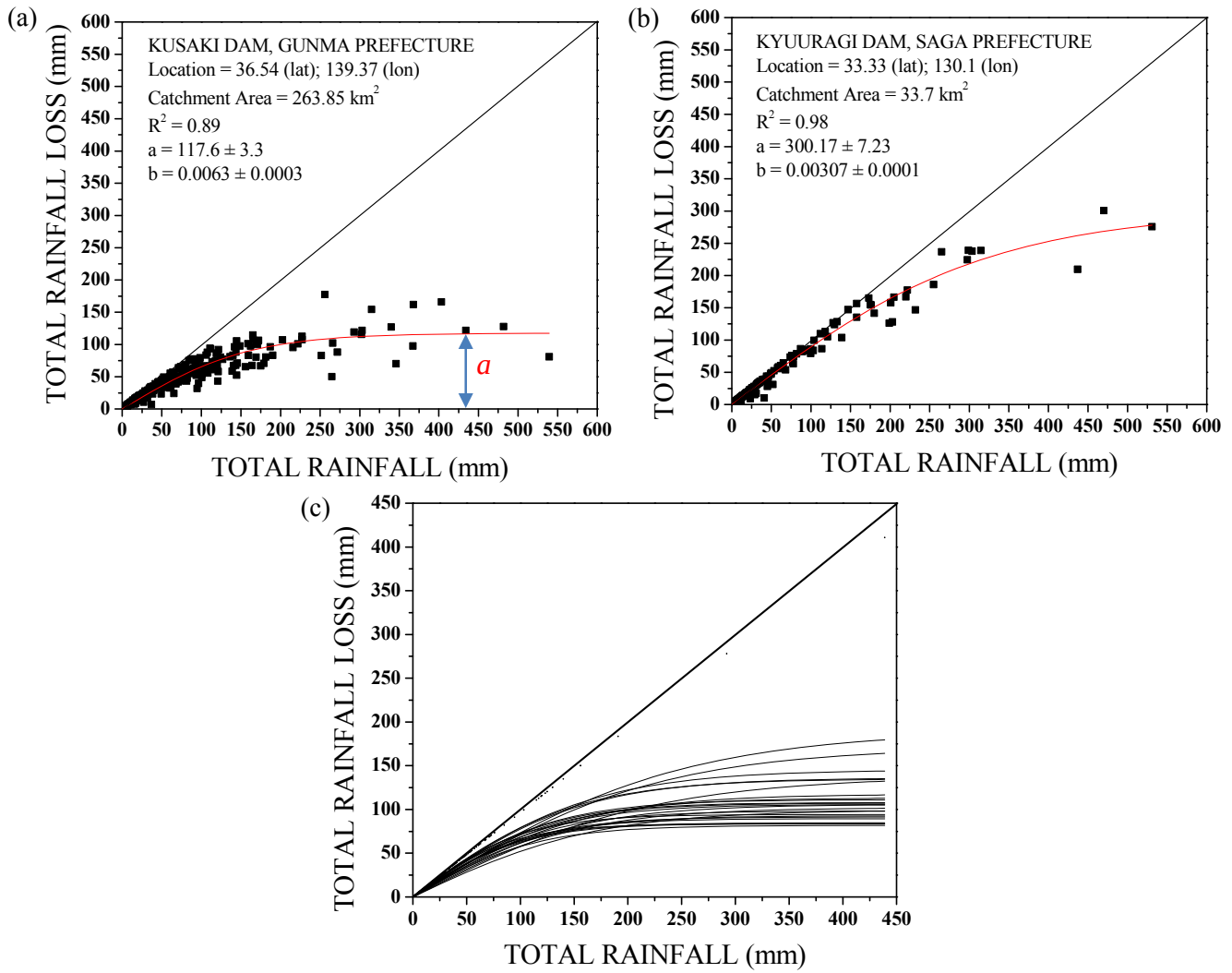


Figure 3.3. Relationship between total rainfall (mm) and total rainfall loss (mm) for the (a) Kusaki dam catchment in Gunma Prefecture, (b) Kyuuragi dam catchment in Saga Prefecture, and (c) 23 catchments in Japan having a constant-stage tanh-type curve.

The values of runoff parameters a and b including their standard deviation values (1σ) that are obtained by applying Eq. (6) for 23 catchments having a constant-stage tanh-type curve like shown in Figure 3.3. (a) are summarized in Table 3.1.

Table 3.1. Summary of values of runoff parameters a and b including their standard deviation values

No.	Dam	Prefecture (Latitude, Longitude)	Island	Catchment Area (km ²)	a	Standard Deviation of a	b	Standard Deviation of b
1	Houheikyou	Hokkaido (42.92, 141.15)	Hokkaido	136.1	91.47	2.9	0.009	0.0004
2	Iwaonai	Hokkaido (44.12, 142.71)	Hokkaido	341.6	89.59	4.7	0.009	0.0006

No.	Dam	Prefecture (Latitude, Longitude)	Island	Catchment Area (km ²)	<i>a</i>	Standard Deviation of <i>a</i>	<i>b</i>	Standard Deviation of <i>b</i>
3	Jyouzankei	Hokkaido (42.98, 141.16)	Hokkaido	103.6	84.75	3.5	0.010	0.0006
4	Kanayama	Hokkaido (43.13, 142.44)	Hokkaido	410.8	108.02	3.5	0.008	0.0004
5	Satunaigawa	Hokkaido (42.59, 142.92)	Hokkaido	116.6	99.27	6.3	0.006	0.0005
6	Tokachi	Hokkaido (43.24, 142.94)	Hokkaido	598.2	136.25	3.6	0.006	0.0002
7	Gosho	Iwate (39.69, 141.03)	Honshu	635.2	92.89	3.2	0.008	0.0004
8	Ikari	Tochigi (36.9, 139.71)	Honshu	271.2	105.93	4.5	0.008	0.0005
9	Kawamata	Tochigi (36.88, 139.52)	Honshu	179.4	107.35	3.4	0.008	0.0004
10	Aimata	Gunma (36.71, 138.89)	Honshu	110.8	113.19	5.5	0.007	0.0005
11	Kusaki	Gunma (36.54, 139.37)	Honshu	263.9	117.57	3.3	0.006	0.0003
12	Simokubo	Gunma (36.13, 139.02)	Honshu	323.7	98.41	4.3	0.008	0.0005
13	Sonohara	Gunma (36.64, 139.18)	Honshu	601.1	135.02	3.1	0.007	0.0002

No.	Dam	Prefecture (Latitude, Longitude)	Island	Catchment Area (km ²)	<i>a</i>	Standard Deviation of <i>a</i>	<i>b</i>	Standard Deviation of <i>b</i>
14	Yagisawa	Gunma (36.91, 139.06)	Honshu	165.5	137.61	6.6	0.004	0.0003
15	Futase	Saitama (35.94, 138.91)	Honshu	170.6	94.30	3.5	0.010	0.0006
16	Koshibu	Nagano (35.61, 137.98)	Honshu	289.6	83.57	4.2	0.011	0.0009
17	Makio	Nagano (35.82, 137.6)	Honshu	307.8	105.77	7.4	0.007	0.0007
18	Miwa	Nagano (35.81, 138.08)	Honshu	311.0	111.26	4.4	0.008	0.0005
19	Yokoyama	Gifu (35.59, 136.46)	Honshu	470.7	145.35	4.5	0.006	0.0003
20	Sintoyone	Aichi (35.13, 137.76)	Honshu	111.4	102.02	6.6	0.006	0.0007
21	Hitokura	Hyogo (34.91, 135.41)	Honshu	115.1	83.91	3.0	0.010	0.0005
22	Nomura	Ehime (33.36, 132.63)	Shikoku	168.0	81.84	4.0	0.009	0.0007
23	Matsubara	Ooita (33.19, 130.99)	Kyushu	491.0	170.92	7.1	0.004	0.0003

Figure 3.3. (a) shows that, in the earlier stage when the rainfall events having smaller total rainfall, the curve is linear, and it then enters the transition stage before finally reaching the constant stage. The physical meaning of this curve can be described as follows: for each rainfall event in the linear stage, the majority portion of rain water become rainfall loss (nearly zero runoff) because the catchment is

not yet saturated (the number of heavy rainfall events are not sufficient to reach saturation conditions). On the contrary, for each rainfall event in the constant stage, major portion of the rain water become direct runoff because the catchment is saturated. Thus, based on the physical interpretations given before to the linear and constant parts of the curves, catchments having a constant-stage tanh-type curve are characterized by a constant stage after the linear stage due to some heavy rainfall events that have small total rainfall loss, whereas catchments having a constant stage demonstrate saturation conditions.

Earlier it has been explained that the plotting result of total rainfall and total rainfall loss by using **Eq. (6)** showed that besides the average values of runoff parameters a and b , the standard deviation of those runoff parameters also could be obtained. In this plotting result between total rainfall and total rainfall loss, the rainfall events varies from small rainfall events having total rainfall in the range of 0-50 mm till big rainfall events that contribute to the saturation condition of a catchment. However, majority of rainfall events are small rainfall events which contributes in the linear part of a tanh curve that eventually affects the constant stage of a tanh curve. Thus, the obtained standard deviation values of runoff parameters a and b at the constant stage are influenced by the small rainfall events in the linear part of the tanh curve.

In this study, runoff parameter a as shown in **Table 3.1** in which the values varies from 81.8 to 170.9 mm, with an average value of 108.5 mm, is interpreted as the height of tanh curve for catchments having a constant-stage tanh-type curve, and is named total rainfall loss under saturation condition because it can be obtained only when the target catchment has ever experienced saturation condition.

The scatter shown in the empirical event values around the calibrated relationship (see **Figure 3.3. (a)**) means that same value of total rainfall may lead to different values of total rainfall loss, due to differences in evaporation that affects the initialization of soil moisture, interception by vegetation cover, or depression storage on the land surface. To give an example, one of the events plotted in **Figure 3.3. (a)** has a total rainfall depth of 367.3 mm and has a total rainfall loss of 97.4 mm, whereas another event of approximately the same total rainfall depth of 368.0 mm produced a total rainfall loss of 161.8 mm. These additional factors cause uncertainty in the quantification of the parameter a , and the standard deviation value of parameter a is being utilized to measure this uncertainty.

By identifying the threshold of total rainfall that causes saturation conditions for each catchment having a constant-stage, the tanh curves can be split in its linear, transition, and constant stages. This threshold was derived by identifying the total rainfall depth where the localized gradient of the tanh curve reaches a value close to zero. This localized gradient is calculated by taking the derivative of **Eq. (6)**:

$$F(R)' = ab \operatorname{sech}^2(bR) \quad (48)$$

where $F(R)'$ is the localized gradient, R is total rainfall required to cause a saturation condition (mm) if **Eq. (6)** is applied to identify when the localized gradient is close to 0. The value of total rainfall causing a saturated condition for each catchment with a constant -stage is presented in **Table 3.2**.

Table 3.2. Summary of seasonal mean precipitation (mm), and total rainfall required to cause saturation conditions (mm) for 23 catchments having a constant-stage tanh-type curve.

Dam	Seasonal mean precipitation from June to October for 10 years (mm)	Total rainfall required to cause saturation conditions (mm)
Houheikyou	722.8	234.6
Iwaonai	699.5	228.4
Jyouzankei	538.7	212.5
Kanayama	706.3	258.8
Satsunaigawa	1168.7	324.3
Tokachi	830.2	327.0
Gosho	1001.7	253.4
Ikari	990.5	269.3
Kawamata	994.4	264.0
Aimata	1044.0	294.1
Kusaki	1276.4	321.6
Simokubo	847.2	253.1
Sonohara	905.3	316.0
Yagisawa	1017.1	431.1
Futase	1062.0	215.8
Koshibu	982.2	194.9
Makio	1311.7	299.9
Miwa	982.7	267.6
Yokoyama	1364.7	346.1
Sintoyone	1477.5	303.5
Hitokura	711.3	213.5
Nomura	965.7	229.3
Matsubara	1391.0	458.6

The minimum, maximum, and average values of these total rainfall depths are 194.9, 458.6, and 283.4 mm, respectively. Given that the minimum value is close to 200 mm, this value is determined to be the threshold of total rainfall that will cause saturation conditions. For the 23 catchments that reach such threshold, the average ratio of total direct runoff to total rainfall is 60.7%, and the average ratio of total rainfall loss to total rainfall is 39.3%. As for the remaining catchments having a non-constant-stage tanh-type curve, the average ratio of total direct runoff to total rainfall is 38.9%, and the average ratio of total rainfall loss to total rainfall is 61.1%.

In the earlier part of this section, it has been described that total rainfall loss is defined as the total

amount of rainfall that does not contribute to direct runoff, and parameter a is defined as total rainfall loss under saturation condition. Thus, parameter a can represent the water holding capacity of a catchment because it reflects the amount of rainfall that can be stored in the catchment and does not contribute to direct runoff when rain occurs. Hence, in this study parameter a in the total rainfall and total rainfall loss relationship is defined as the parameter to explain about the potential catchment storage. Therefore, if a catchment has a bigger value of a , it means the catchment has a larger capacity to store water.

Among 23 catchments having a constant-stage tanh-type curve, there are 13 catchments having value of a more than 100 mm where 1 catchment is located on Kyushu Island, 2 catchments are located on Hokkaido Island, and 10 catchments are located on Honshu Island (see **Table 3.1**). Therefore, majority of catchments with a higher value of a are located on Honshu Island. The maximum value of a , 170.9 mm, was found in the Matsubara Dam catchment in Ooita Prefecture on Kyushu Island (33.19°N, 130.99°E), whereas the minimum value, 81.8 mm, was found in the Nomura Dam catchment in Ehime Prefecture on Shikoku Island (33.36°N, 132.63°E) (see **Figure 2.1 and Table 3.1**).

3.2 STANDARD DEVIATION OF PARAMETER a

It has been explained in the methodology that the obtained values of a and b for catchments having constant-stage tanh type curve as shown in **Table 3.1** are used as the input data to calculate outflow contribution rate by using **Eq. (10)**. Then the outflow contribution rate is used to calculate the effective rainfall intensity by using **Eq. (11)**. The obtained effective rainfall intensity is then used as the input data to simulate runoff height by using **Eq. (31)**. The value of outflow contribution rate reflects the initial catchment condition, where if the outflow contribution rate is equal to 1, it means that the initially catchment condition is saturated, hence that there is no infiltration and all rainfall becomes direct runoff.

It can be seen from **Eq. (10)** that the bigger values of a and b results in smaller values of outflow contribution rate. It means that catchments with bigger values of a and b having higher infiltration and less direct runoff. The higher infiltration occurred when initially catchment condition was dry, so majority of portion of rainfall can infiltrate into the ground. On the contrary, smaller values of a and b results in bigger values of outflow contribution rate, which means catchments with smaller values of a and b having lower infiltration and more direct runoff. The lower infiltration occurred when initially catchment condition was wet, so catchment has already saturated, thus majority of portion of rainfall become direct runoff.

It has been discussed that the plotting of total rainfall and total rainfall loss relationship yields runoff parameters a and b values, and also their standard deviation values ($\pm 1\sigma$) that reflect the uncertainty of obtained parameters a and b (see **Figure 3.4 (a)**).

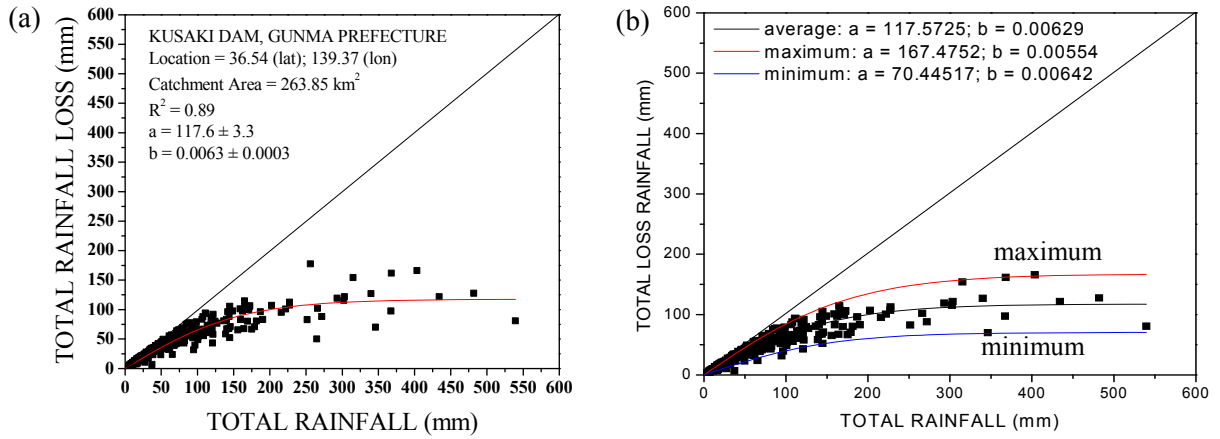


Figure 3.4. (a) Standard deviation values to explain about initial water amount, (b) Maximum and minimum values of runoff parameters to explain about extreme cases.

By adding the runoff parameters a and b with the obtained values of 1σ ($a + 1\sigma$ of a and $b + 1\sigma$ of b) shown in **Table 3.1**, the values of a and b become bigger, and results in higher infiltration, which means initially catchment was dry. On the contrary, by subtracting the runoff parameters a and b with the obtained values of 1σ ($a - 1\sigma$ of a and $b - 1\sigma$ of b), the values of a and b become smaller, and results in smaller infiltration, which means initially catchment was wet. Thus, the negative value of standard deviation can represent the wet initial soil moisture condition, whereas the positive value of standard deviation can represent the dry initial soil moisture condition. Aside from standard deviation, the maximum and minimum values are shown to demonstrate the influence extreme dry or wet initial soil-moisture conditions (when the catchment is initially very dry or wet before rain occurs) (see **Figure 3.4 (b)**). The geographic distribution of standard deviation of parameter a is shown in **Figure 3.5**.

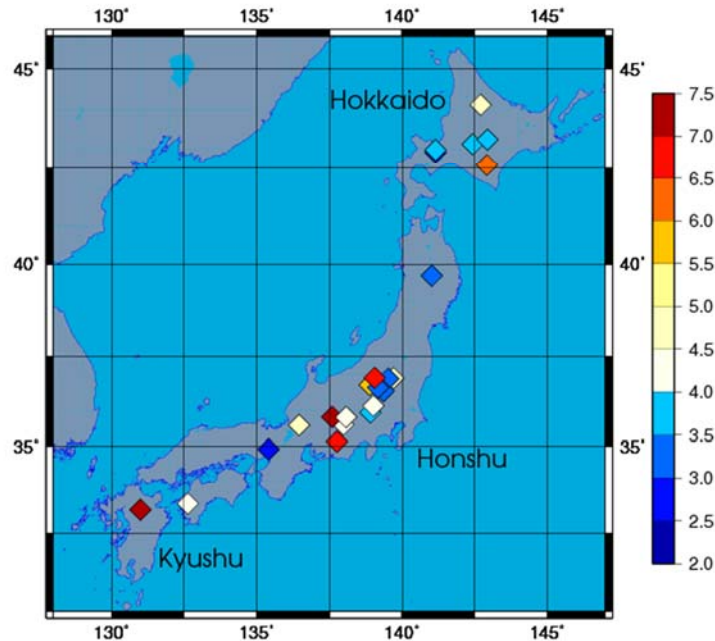


Figure 3.5. Geographic distribution of standard deviation of total rainfall loss under saturation conditions for 23 catchments having a constant-stage tanh-type curve.

The standard deviation of parameter a varies from 2.9 mm to 7.4 mm, with an average value of 4.5 mm. The maximum value of standard deviation of parameter a , 7.4 mm, was found in the Makio dam catchment in Nagano Prefecture on Honshu Island (35.82°N, 137.60°E), whereas the minimum value, 2.9 mm, was found in the Houheikyou dam catchment in Hokkaido Prefecture on Hokkaido Island (42.92°N, 141.15°E) (see **Figure 3.5**).

3.3 SUMMARY

The new findings found on this chapter can be summarized as follows:

- Among 36 catchments in Japan that having continuous dataset for the target period in this study i.e. June-October at least from 2002 till 2011, results showed that majority of rainfall events having total rainfall in the range of 0-50 mm. It means that majority of rainfall events in Japan occurred with small total rainfall.
- By plotting total rainfall and total rainfall loss relationship for those 36 catchments, results showed that 23 catchments having a constant-stage tanh type curve, whereas 13 catchments having a non-constant stage tanh type curve. A constant-stage tanh type curve is characterized by linear stage in the beginning of the curve, and it then enters the transition stage before finally reaching the constant stage. The constant stage demonstrate saturation conditions of the catchment. Runoff parameters a and b only can be obtained for catchments having a constant-stage tanh type, thus the effective rainfall intensity as the input data to simulate runoff can be calculated only for those 23 catchments. These results also indicate that majority of catchments in Japan have not experienced saturation condition yet due to lacking of number of heavy rainfall events with small total rainfall loss to cause saturation condition.
- In this study, runoff parameter a in the total rainfall and total rainfall loss relationship is interpreted as the height of tanh curve, and is proposed as the parameter to estimate potential catchment storage. This runoff parameter a is named total rainfall loss under saturation condition because it can be obtained only when a catchment has ever experienced saturation condition.
- Among those 23 catchments having constant-stage tanh type, there are 13 catchments having value of runoff parameter a more than 100 mm, which a bigger value of runoff parameter a indicates greater capacity of catchment storage. Among those 13 catchments, 10 catchments are located on Honshu Island. Thus, majority of catchments having greater capacity to store water during a rain event are located on Honshu Island.
- The runoff parameter a consists of standard deviation ($\pm 1\sigma$) value. The negative value of standard deviation can represent the wet initial soil moisture condition, whereas the positive value of standard deviation can represent the dry initial soil moisture condition. Thus, the standard deviation value of runoff parameter a is used to represent the uncertainty associated with water holding capacity to know how much water initially stored in the ground before the target rainfall event occurred.

Chapter 4. PROPOSED ESTIMATION METHOD OF POTENTIAL CATCHMENT STORAGE FOR CATCHMENTS THAT NEVER EXPERIENCE SATURATION CONDITIONS

The content of this chapter is submitted to the Journal of Hydrology Research.

4.1 ESTIMATION OF RUNOFF PARAMETERS (a and b) AT THE CONSTANT STAGE

In the previous sections, it was explained that runoff parameters in the total rainfall – total rainfall loss relationship can be used to estimate effective rainfall intensity only for catchments having a constant-stage tanh-type curve, which are characterized by the presence of heavy rainfall events with small total rainfall loss. This raises the question as to how the total rainfall losses can be assessed for catchments that never experienced saturation conditions. Two methods namely localized gradient method and inverse method are presented below. They are tested/evaluated based on the 23 catchments that have constant-stage tanh-type curve, but after selecting a limited set of rainfall events. When such limited set is considered, less events are expected to cause saturation conditions. **Figure 4.1** shows that the majority of the events have total rainfall depths in the range 0 – 50 mm, and that the frequency gradually decreases with each increment in total rainfall. This explains why several catchments did not experience saturation conditions yet, and why consideration of a limited set of events reduces the likelihood that such saturation conditions are reached.

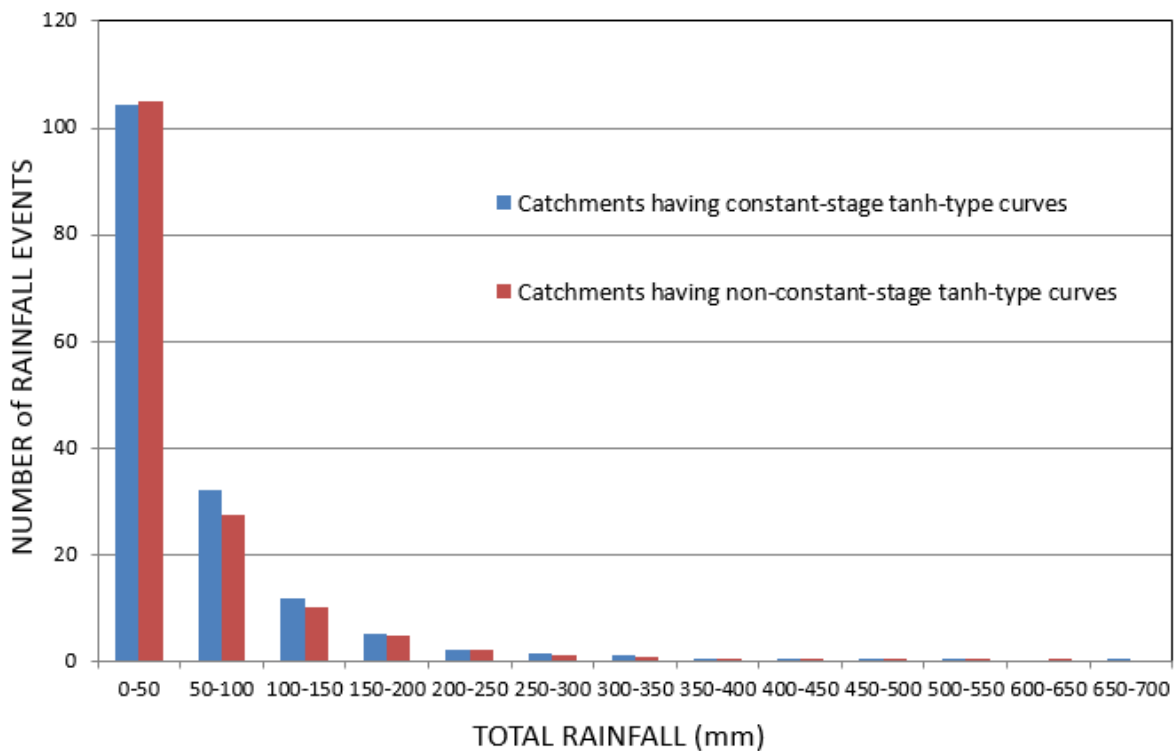


Figure 4.1. Average frequency of rainfall events for each range of total rainfall. The blue bars denote catchments having a constant-stage tanh-type curve, and the red bars denote catchments having a non-constant-stage tanh-type curve.

4.1.1 LOCALIZED GRADIENT METHOD

The first method is based on localized gradient calculations and on analyzing similarity with the conditions at the catchments with constant-stage tanh-type curve. Because section 3.1 explained that total rainfall of about 200 mm causes saturation conditions, the different ranges of total rainfall are considered to be 0–100, 0–150, 0–175, and 0–200 mm. For each of these ranges, the new values of runoff parameters a and b can be obtained using the same method as before, hence after plotting the total rainfall – total rainfall loss relation using **Eq. (6)**. Then, the new values of a and b for different ranges of total rainfall are plotted against the values of a and b at the constant stage for the 23 catchments having a constant-stage tanh-type curve. The correlation between the values of a at the constant stage and the new values of a for the different total rainfall classes shows that rainfall events in the class of 0–200 mm have the highest correlation among all classes (correlation coefficient of 0.87). The correlation between the values of b at the constant stage and the new values of b for the different total rainfall classes moreover shows that the total rainfall class of 0–200 mm has the highest correlation among all rainfall classes (correlation coefficient of 0.74). Thus, the new values of a and b obtained based on **Eq. (6)** for the total rainfall class of 0–200 mm are adopted for calculating the localized gradient.

The next step is to calculate the localized gradient based on selected total rainfall values for different target rainfalls, namely 75, 100, 125, 150, 175, and 200 mm. After obtaining the new values of a and b , the localized gradient for a specific target rainfall can be calculated by adopting **Eq. (48)** after $F(R)'$ is renamed as y and R is renamed as x :

$$y = absech^2(bx) \quad (49)$$

where y is the localized gradient, x is the target rainfall (rainfall that does not cause saturation conditions) (mm), and a and b are the new values of a and b depending on the range of total rainfall, i.e. total rainfall that does not cause saturation conditions. Thus, the localized gradient for each target rainfall can be calculated using **Eq. (49)**.

For each target rainfall, based on the total rainfall class of 0–200 mm, the obtained localized gradient for the 23 catchments with constant-stage tanh-type curve are plotted against the values of a at the constant stage. The correlation coefficients between the values of a at the constant stage and the localized gradient of the 23 catchments with constant-stage tanh-type curve for each target rainfall amount reveal that at a total rainfall of 0–200 mm, target rainfall of 200 mm has the highest correlation with the values of a at the constant stage (correlation coefficient of 0.90). Hence, the total rainfall class 0–200 mm is selected for estimating the value of a and b at the constant stage for a target rainfall of 200 mm.

The empirical equation to estimate the value of a at the constant stage is as follows:

$$y = 0.0031ea - 0.2221 \quad (50)$$

or:

$$ea = \frac{y + 0.2221}{0.0031} \quad (51)$$

where ea is estimated a at the constant stage, and y is the obtained localized gradient.

To estimate the value of b at the constant stage, the new values of a and b for the total rainfall class of 0–200 mm across the 23 catchments with a constant-stage tanh-type curve are plotted to obtain the following empirical equation:

$$eb = 0.6187ea^{-0.936} \quad (52)$$

where eb is the estimated value of b at the constant stage, and ea is the estimated value of a at the constant stage calculated using **Eq. (51)**. The correlation coefficient between the new values of a and b for the total rainfall class of 0–200 mm is 0.92.

4.1.2 INVERSE METHOD

For the localized gradient method, the estimation of runoff parameters at the constant stage is based on the data of the 23 catchments that have a constant-stage tanh-type curve. For application purposes, it is more practical to estimate the runoff parameters at the constant stage using information from the target catchment only. Hence, another method, called the inverse method, is proposed. The main idea of this method is to use the observed discharge, which usually is the target output, as the input data.

Previously, it has been explained that the majority of rainfall events have total rainfall in the range of 0–50 mm, whereas total rainfall of 200 mm has been determined as the threshold causing saturated conditions. Hence, the threshold for selecting a limited data set of total rainfall that does not cause saturation conditions but that can contribute to estimating the runoff parameters at the constant stage is in the range of 50–200 mm. More specifically, the following total rainfall ranges are chosen: 50–190, 50–180, 50–170, 50–160, 50–155, 50–140, and 50–130 mm. To test the method, for each range, a limited set of rainfall events are chosen randomly for analysis.

The total runoff of the rainfall events is calculated by dividing the total observed discharge by the area of the target catchment. After calculating the total runoff derivative $\left(\frac{dq_s}{dt}\right)$, the effective

rainfall intensity is calculated by rearranging **Eq. (31)** as follows:

$$r_e(t) = \left[\frac{dq_*(t)}{dt} / a_0 q_*(t)^\beta \right] + q_*(t) \quad (53)$$

where after the outflow contribution rate is calculated by rearranging **Eq. (11)**:

$$ocr(t) = r_e(t)/r(t) \quad (54)$$

The next step is to apply **Eq. (10)** to the rainfall events to estimate the runoff parameters a and b . The two unknowns, a and b , are obtained by applying a numerical nonlinear global optimization algorithm with constraints. The constraint applied here is that the a value must be positive, because a represents the total rainfall loss under saturation conditions. By considering several combinations of randomly selected rainfall events for each total rainfall range, different values of a and b are obtained, and the ensemble averages used as the estimates of a and b at the constant stage.

4.2 CASE STUDY

The Kusaki dam catchment in Gunma Prefecture is selected as the case study because it has the longest data set (30 years). The maximum total rainfall during the full 30 years was 539.5 mm. Of the 250 rainfall events that have occurred during this period, 22 events have total rainfall higher than 200 mm. By plotting the tanh function using **Eq. (6)** for the whole target period, the obtained runoff parameters at the constant stage for a is 117.6, and that for b is 0.0063.

These estimates are compared with the two methods proposed to estimate the values of a and b at the constant stage by using smaller rainfall events (limited data set). For the localized gradient method, the estimated values calculated using **Eq. (51)** and **Eq. (52)** are 111.7 mm for a and 0.0075 for b . For the inverse method, the estimated values of a and b for different ranges of total rainfall are presented in **Table 4.1**.

Table 4.1. Parameter values for a as total rainfall loss under saturation conditions (mm) and b based on the inverse method for different ranges of total rainfall.

Threshold	a	Δa (%)	b	Δb (%)
50-190 mm	119.7	1.8	0.0062	1.6
50-180 mm	119.2	1.4	0.0059	6.3
50-170 mm	126.0	7.1	0.0052	17.5
50-160 mm	120.5	2.5	0.0059	6.3
50-155 mm	141.8	20.6	0.0044	30.1
50-140 mm	104.4	11.2	0.0067	6.3
50-130 mm	99.0	15.7	0.0064	1.6
Average (constant)	117.6	0	0.0063	0

The estimated values of a and b at the constant stage using each method are then used to simulate discharge for 26 rainfall events having total rainfall ranging from 45.0 to 364.0 mm. For the inverse method, several values of a and b are possible based on different ranges of total rainfall. The selected values of a and b can be obtained by doing validation. If the discrepancies of peak discharge and time to peak are acceptable, then those a and b values are chosen as the estimated runoff parameters at the constant stage. In this case, the values of a and b obtained by using total rainfall of 50–160 mm are adequate for use in simulating discharge. After estimating a and b at the constant stage, effective rainfall intensity is calculated using Eq. (53). Then, the discharge is simulated using Eq. (31).

For validation purposes, the simulated discharges are plotted against the observed discharges. Figure 4.2 shows this comparison for a target rainfall.

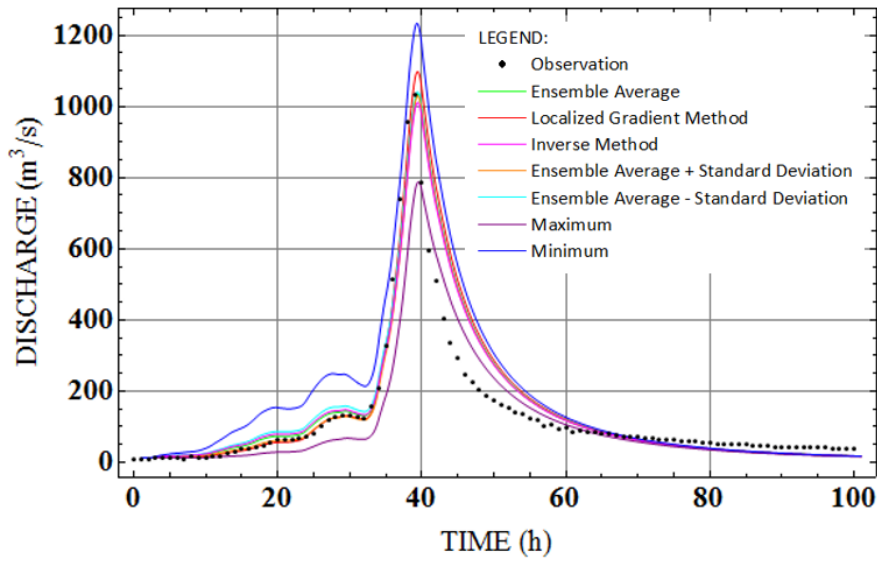


Figure 4.2. Comparison between simulated and observed discharges of the target rainfall for the Kusaki dam catchment in Gunma Prefecture, Japan.

Two important parameters for describing peak or flood flows, namely the peak discharge (Q_p) and the time to peak (t_p), are quantified. The discrepancies between the simulated Q_p and t_p and their observed values are calculated for each rainfall event. The discrepancies in peak discharge (ΔQ_p) are calculated in terms of percentages, and those in time to peak (Δt_p) are calculated in hours. Table 4.2 reports these ΔQ_p and Δt_p for the target rainfall. The ensemble averages of ΔQ_p and Δt_p based on all 26 rainfall events are presented in Figure 4.3. The ensemble average of ΔQ_p equals 7.6%, and the ensemble average of Δt_p is 1.2 hours. For discharge simulated by the inverse method, the ensemble average of ΔQ_p is 8.5%, and the ensemble average of Δt_p is 1.3 hours. Based on the localized gradient method, the ensemble average of ΔQ_p is 9.7%, and the ensemble average of Δt_p is 1.3 hours. The error bars represent the standard deviations calculated based on the differences of the results for all events versus the ensemble average. These standard deviations represent the influence of the initialization of the soil moisture condition. Next to the ensemble mean and standard deviation, the maximum and minimum values are shown to demonstrate the influence extreme dry or wet initial soil-moisture conditions (when the catchment is initially very dry or wet before rain occurs).

The ensemble average plus/minus its standard deviation give a value of 10.4% /13.1% for ΔQ_p , and 1.2/1.2 hours for Δt_p . The maximum/minimum values of ΔQ_p are 49.0%/80.5% and 1.5/1.5 hours for Δt_p .

Table 4.2. Results of ΔQ_p and Δt_p of the target rainfall for Kusaki dam catchment in Gunma Prefecture, Japan.

Case	Peak Discharge, Q_p (m^3/s)	Time to peak, t_p (h)	ΔQ_p (m^3/s)	Δt_p (h)
Ensemble Average	1034.77	39.42	4.36	0.42
Ensemble Average + Standard Deviation	1029.29	39.42	1.12	0.42
Ensemble Average - Standard Deviation	1040.75	39.41	10.34	0.41
Localized Gradient Method	1097.64	39.40	67.23	0.40
Inverse Method	1010.73	39.42	19.68	0.42
Maximum	787.79	39.52	242.62	0.52
Minimum	1233.31	39.33	202.9	0.33
Observation	1030.41	39.00	0	0

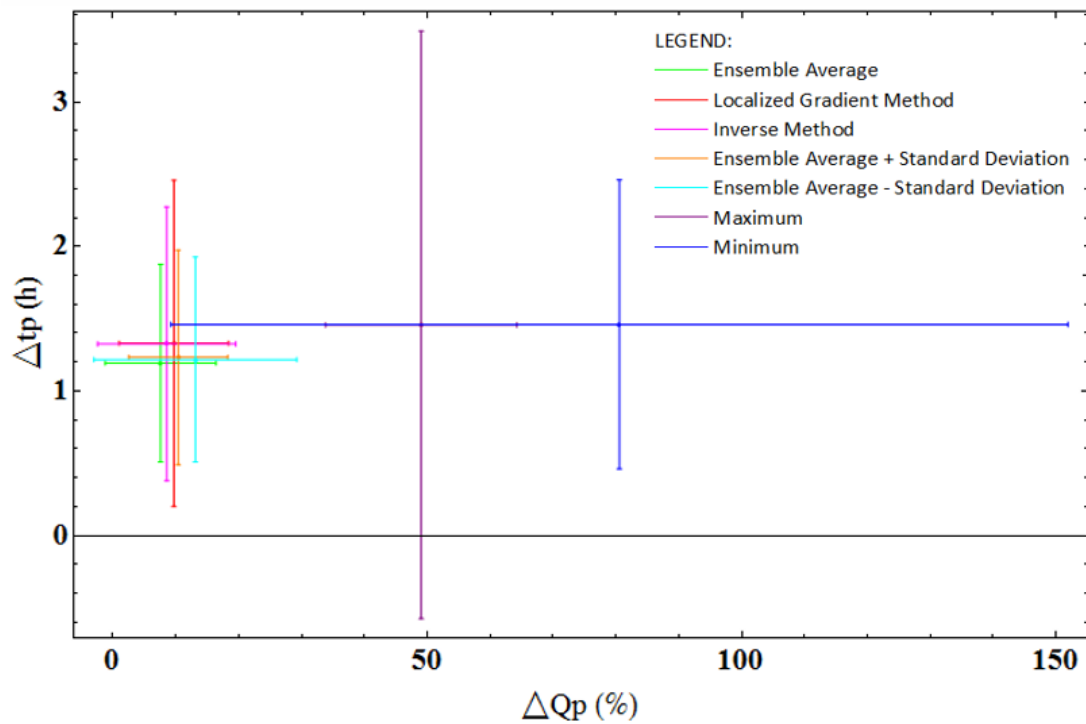


Figure 4.3. Relationship between average values of ΔQ_p (%) and Δt_p (h) with the corresponding standard deviation values for the Kusaki dam catchment in Gunma Prefecture, Japan, based on an ensemble of 26 rainfall events.

4.3 SUMMARY

The new findings found on this chapter can be summarized as follows:

- The results from Chapter 3 showed that many catchments in Japan have never experienced saturation condition, thus runoff parameters a and b to estimate effective rainfall intensity to simulate runoff cannot be obtained for those catchments. Thus, in this chapter, two different methods namely localized gradient method and inverse method are proposed to estimate runoff parameters a and b for those catchments having a non-constant stage tanh-type curve. Those two methods have different characteristics, which localized gradient method is an empirical based method, whereas inverse method is a physical based method, but both methods have similarities in the usage of rainfall events that do not cause saturation condition.
- It is concluded that each of the two proposed methods has advantages and disadvantages. The inverse method does not require information from so many catchments, uses fewer rainfall events, and does not require to calibrate a relationship between the a and b parameters. Nevertheless, to define which range of rainfall event is acceptable for use, trial and error is required. For the localized gradient method, the range of rainfall events to be used is fixed at 0–200 mm, but it requires information from many catchments with a constant-stage tanh-type curve to produce empirical equations for estimating the runoff parameters at the constant stage.
- The obtained results show that smaller rainfall events can be used to estimate the runoff parameters at the constant stage. The two methods show comparable results, but the inverse method shows that smaller rainfall events up to 160 mm are adequate for identifying the runoff model parameters.
- The obtained estimated runoff parameters a and b are then used to simulate runoff for 26 rainfall events occurred at Kusaki dam catchment. For validation purpose, the discrepancy of peak discharge and time to peak between simulated runoff and observed runoff are quantified. Results showed that the discrepancy of peak discharge and time to peak of localized method and inverse method are small (comparable to the simulated runoff using the average values of a and b as the ideal case), thus the proposed methods are reliable to estimate runoff parameters a and b for catchments having a non-constant stage tanh-type curve.

Chapter 5. THE MINIMUM TOTAL RAINFALL REQUIRED TO GENERATE DIRECT RUNOFF

This chapter is written based on a published paper:

Intan Supraba and Tomohito J. Yamada, “Potential water storage capacity of mountainous catchments based on catchment characteristics”, *Journal of Japan Society of Civil Engineers, Ser.B1 (Hydraulic Engineering)*, Vol. 71, No. 4, I_151-I_156, 2015.

The content of this chapter has been presented in the following conference:

Intan Supraba and Tomohito J. Yamada, “Potential Water Storage Capacity of Mountainous Catchments Based On Catchment Characteristics”, in The 59th Annual Conference on Hydraulic Engineering of Japan Society of Civil Engineers, 9th-12th March 2015, Waseda University, Japan.

5.1 THRESHOLD OF MINIMUM TOTAL RAINFALL REQUIRED TO GENERATE DIRECT RUNOFF

Section 3.1 discussed that runoff parameter a can represent the actual water holding capacity as the potential catchment storage. The parameter a is named total rainfall loss under saturation condition because it can be obtained only when the catchment previously ever experienced saturated condition due to some heavy rainfall events with small total rainfall loss. The saturation condition is indicated by constant stage in the tanh curve (see **Figure 3.3. (a)**). This chapter is to discuss the minimum total rainfall required to generate direct runoff. The threshold of minimum total rainfall required to generate direct runoff is defined as the amount of rainfall that almost does not contribute to direct runoff at all, or nearly zero direct runoff, when it rains. The threshold is determined as at least 95% of total rainfall becomes total rainfall loss such as infiltrates into the ground, and a maximum of 5% of total rainfall becomes total direct runoff.

Opposite from parameter a that is obtained at the constant stage of a tanh curve, the threshold of minimum total rainfall required to generate direct runoff is found at the linear stage of a tanh curve, and indeed, it is plotted close to the 1 to 1 plot between total rainfall and total rainfall loss. The geographical distribution of minimum total rainfall required to generate direct runoff is presented in **Figure 5.1**. The minimum total rainfall required to generate direct runoff varies from 7.9 mm to 173.2 mm, with an average value of 51.3 mm. The higher minimum value indicates a higher capacity of catchment storage. The highest and lowest values of the minimum total rainfall required to generate direct runoff were found in the Satsunaigawa dam catchment in Hokkaido Prefecture (42.59°N, 142.92°E), and the Kyuuragi dam catchment in Saga Prefecture (33.33°N, 130.1°E), respectively (see **Figure 5.1**).

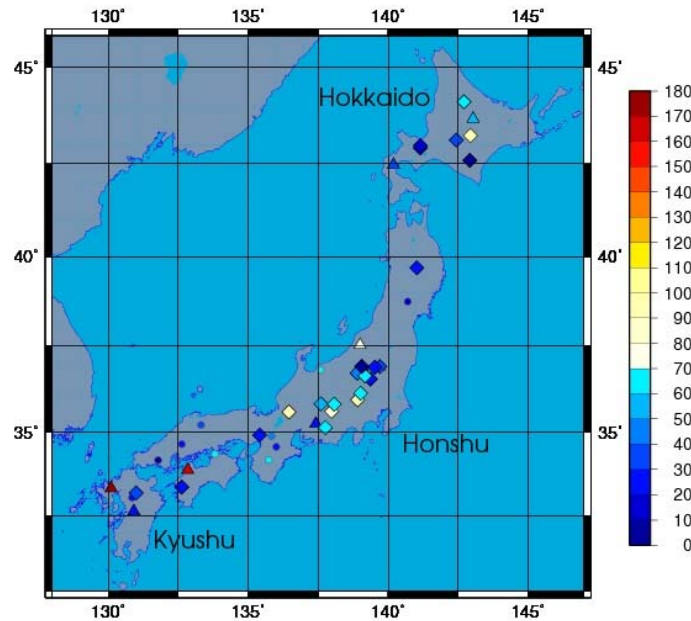


Figure 5.1. Geographic distribution of 47 catchments. Diamonds denote catchments having a constant-stage tanh-type curve, the colored bar indicates the value of the minimum total rainfall required to generate direct runoff (mm), triangles denote catchments having a non-constant-stage tanh-type curve, and circles denote other catchments that lacked complete datasets.

5.2 VARIATION OF POTENTIAL CATCHMENT STORAGE

From section 3.1 and section 5.1, results show that the values of total rainfall loss under saturation condition and minimum total rainfall required to generate direct runoff as important runoff parameters for assessing the potential water storage capacity of a catchment varies. The variation is explained by catchment characteristics (soil types, rock types, terrain types, and vegetation types), and catchment morphometric parameters.

Analysis of water holding capacity for catchment areas ranging from 1 km² to 10 km² in size based on rock types which were classified into three groups namely Quaternary volcanic rocks, Tertiary granitic rocks, and Paleozoic rocks types showed that catchments consisting of Quaternary volcanic rocks have the largest water holding capacity (Musiake, 1978; Musiake et al., 1981). Another analysis of water holding capacity and catchment storage of 52 catchments in Japan with catchment areas ranging from 0.01 to 100 km² has been performed to determine the function of the forest in headwater conservation. The catchment storage capacity was estimated by rainfall amounts ranging from 50 to 250 mm, and mainly depended upon the surface geology and soil type. The largest catchment storage was identified in catchments covered by granite and volcanic ash (Fujieda, 2007). However, after obtaining the information of soil types, rock types, terrain types, and vegetation types, the classification method is only available for the rock types. For that reason and because of many previous studies focused on rock types to analyze the catchment storage capacity, thus in this study only rock types are analyzed to explain the variation of catchment storage capacity.

Catchment morphometric analysis is used to measure the configuration of the earth's surface, and the shape and dimension of its landforms (Clarke, 1966). The morphometric parameters are classified into three aspects; namely, linear (stream length, stream order, bifurcation ratio), areal (drainage density, elongation ratio, catchment width, overland flow length), and relief (average slope, relative relief) (Clarke, 1966; Nongkynrih and Husain, 2011; Seyhan, 1976). The prominent morphometric parameters for analysis of flash flood severity are those related to basin shape and topography (Suseno, 2013). Thus, drainage density, catchment width, longest drainage length, and elongation ratio are selected as parameters for this study.

5.2.1 CATCHMENT MORPHOMETRIC PARAMETERS

The drainage density mesh metadata file was created in the year 1979 with resolution 1:25,000-1:49,999. After creating the catchment boundary consists of river network by using a Digital Elevation Model (DEM), the information of catchment morphometric parameters can be obtained by overlapping the catchment boundary and the river network. The description of catchment morphometric parameters (Langbein, 1947):

(a) Drainage density (D)

Drainage density is a measure of channel spacing to determine how well a catchment is drained by

stream channels. The sum of the lengths of all drainages in a catchment divided by the catchment area results in drainage density.

$$D = L/A \quad (55)$$

where D is drainage density (m/km²), L is the summation of total drainage length (m), A is the catchment area (km²).

(b) Longest drainage length (L_b)

Longest drainage length is the length of the longest actual stream within this catchment, or the length of main channel. The unit is in km.

(c) Elongation ratio (R_e)

Elongation ratio is the ratio between the diameter of the circle of the same area as the drainage basin and the maximum length of the basin.

$$R_e = 1.129 \frac{\sqrt{A}}{L_b} \quad (56)$$

where R_e is elongation ratio. Elongation ratio measures the shape of the catchment related to the length of main channel. $R_e = 1$, the shape of catchment is circular. $R_e < 1$, the shape of catchment is ellipse with main channel tends to be parallel with the major axis. $R_e > 1$, the shape of catchment is ellipse with main channel tends to be parallel with the minor axis.

(d) Catchment width (W)

The catchment width can be calculated by dividing the catchment area by the longest drainage length.

$$W = A/L_b \quad (57)$$

where W is the catchment width (km).

The relationships between catchment morphometric parameters and the runoff parameters (i.e., total rainfall loss under saturation condition and minimum total rainfall required to generate direct runoff) for 23 catchments having a constant-stage tanh-type curve are analyzed based on correlation coefficient analysis. The area of those catchments varies from 103.6 km² to 635.2 km². The catchments are classified by their areas using a cluster analysis that represent the approximate groupings of the catchment areas based on distance or dissimilarity function. Identical catchment areas have zero distance or dissimilarity, and all of the others have positive distance or dissimilarity. Based on the cluster analysis, the 23 catchments could be classified into three groups; namely, catchments with an area of 100–200 km², catchments covering an area of 200–450 km², and catchment with an area of 450–650 km².

The correlation coefficients between the catchment morphometric parameters and runoff parameters for different groups of catchment areas are presented in **Table 5.1 (a)** and **Table 5.1 (b)**.

Results show that the runoff parameters are closely correlated with the drainage density, elongation ratio, and catchment width for catchments with larger areas (450–650 km²). In catchments with smaller areas (100–450 km²), all of the morphometric parameters are poorly correlated with those runoff parameters.

Total rainfall loss under saturation condition and minimum total rainfall required to generate direct runoff are closely related to the permeable areas of the catchment. Dirac's delta function in **Eq. (9)** represents the fraction of impermeable area in the catchment (Yamada, 2003; Kure and Yamada, 2004). The correlation coefficients between the morphometric parameters of the catchment and Dirac's delta function are presented in **Table 5.1 (c)**.

Table 5.1. Correlation coefficient between catchment morphometric parameters and a) total rainfall loss under saturation condition, b) minimum total rainfall required to generate direct runoff, and c) Dirac's delta function.

(a)	Catchment morphometric parameters	Correlation coefficient for catchment area		
		100-200 km ²	200-450 km ²	450-650 km ²
	Drainage density (m/km ²)	0.13	0.02	0.99
	Longest drainage length (km)	0.11	0.02	0.36
	Elongation ratio	0.19	0.04	0.98
	Width of watershed (km)	0.15	0.03	0.92

(b)	Catchment morphometric parameters	Correlation coefficient for catchment area		
		100-200 km ²	200-450 km ²	450-650 km ²
	Drainage density (m/km ²)	0.05	0.14	0.90
	Longest drainage length (km)	0.02	0.07	0.33
	Elongation ratio	0.03	0.04	0.93
	Width of watershed (km)	0.03	0.05	0.89

(c)	Catchment morphometric parameters	Correlation coefficient for catchment area		
		100-200 km ²	200-450 km ²	450-650 km ²
	Drainage density (m/km ²)	0.00	0.27	0.86
	Longest drainage length (km)	0.12	0.06	0.51
	Elongation ratio	0.22	0.19	0.87
	Width of watershed (km)	0.15	0.23	0.71

Catchment morphometric analysis has an important role to understand the geo-hydrological behavior of a catchment. The catchment metamorphic parameters and runoff parameters relationships show higher correlation coefficients for the larger catchment areas (450–650 km²) with regard to drainage density, elongation ratio, and catchment width (see **Table 5.1 (a)** and **Table 5.1 (b)**). Higher

drainage density, elongation ratio, and catchment width result in less minimum total rainfall required to generate direct runoff, and lower total rainfall loss under saturation condition. The drainage density can provide a quantitative measure of the average length of stream channels in the entire catchment (Nongkynrih and Husain, 2011; Horton, 1932); a higher drainage density is equivalent to a longer length of stream channels, so more rainwater can be drained out to the outlet. Thus, less rainwater infiltrates into the ground, resulting in smaller values of both, total rainfall loss under saturation condition and minimum total rainfall required to generate direct runoff. The elongation ratio represents the shape of the catchment, with a higher elongation ratio indicating that the major part of catchment is of high relief (Nongkynrih and Husain, 2011). A catchment with high relief can quickly drain rainwater to the outlet; thus, a higher elongation ratio causes higher direct runoff and less infiltration. A wider catchment has a higher drainage density; thus, a wider catchment gives rise to higher direct runoff and less infiltration.

The relationships between Dirac's delta function and catchment parameters show that higher values of drainage density, elongation ratio, and catchment width result in a larger value for Dirac's delta function (see **Table 5.1 (c)**), which in turn, is equivalent to a larger impermeable area in the catchment, resulting in higher direct runoff. In addition, higher drainage density, elongation ratio, and catchment width result in higher direct runoff as explained previously, and this justifies the linear relationship between those three catchment morphometric parameters and Dirac's delta function.

Previous studies have suggested that, for smaller catchment areas, the influence of river channel networks in the catchment can be neglected (Kure and Yamada, 2006; Tachikawa et al., 2003). This explains why catchment morphometric parameters are closely correlated with runoff parameters for larger catchments areas only.

5.2.2 CATCHMENT CHARACTERISTICS

The raw data of rock types are obtained from the MLIT database. After obtaining the data of rock types, the next step is to classify them in order to analyze the correlation between those catchment characteristics and catchment storage capacity. The rock types are classified by using the established method i.e. 3 rock classes (igneous, metamorphic, and sedimentary) (Fetter, 2000; Pipkin et al., 2005; Weight, 2008). The classification of 36 rock types into 3 rock classes is listed in **Table 5.2**.

Table 5.2. Rock classification for some catchments in Japan.

No.	Dam Catchment	Catchment Area (km ²)	Rock Types	Area (%)	Area (km ²)	Rock Classes
1	Houheikyuu	136.1	Andesitic rocks	90.9	123.7	Igneous
			Volcanic breccia, Tuff breccia	7.3	10.0	Sedimentary
			Sandstone	1.3	1.8	Sedimentary
			Rock tuff	0.5	0.6	Igneous

No.	Dam Catchment	Catchment Area (km ²)	Rock Types	Area (%)	Area (km ²)	Rock Classes
2	Iwaonai	341.58	Andesitic rocks	37.2	127.1	Igneous
			Diabase rocks	4.9	16.6	Igneous
			Gabbro rocks	2.1	7.1	Igneous
			Granitic rocks	0.8	2.8	Igneous
			Slate	17.5	59.8	Metamorphic
			Hornfels	2.4	8.0	Metamorphic
			Conglomerate	19.1	65.3	Sedimentary
			Gravel, Sand, Clay	6.5	22.4	Sedimentary
			Sandstone	5.5	18.8	Sedimentary
			Alteration Sandstone, Mudstone	2.2	7.4	Sedimentary
			Volcanic breccia, Tuff breccia	0.7	2.5	Sedimentary
			Mudstone	0.6	1.9	Sedimentary
			Gravel, Sand	0.5	1.8	Sedimentary
			Sandstone, Conglomerate	0.0	0.0	Sedimentary
3	Izarigawa	113.25	Pumice	55.6	63.0	Igneous
			Andesitic rocks	24.2	27.4	Igneous
			Mudstone	12.6	14.3	Sedimentary
			Sandstone	5.3	6.0	Sedimentary
			Gravel, Sand, Clay	1.7	2.0	Sedimentary
			Clastic	0.4	0.5	Sedimentary
			Sand, Gravel, Clay	0.1	0.1	Sedimentary
4	Jyouzankei	103.59	Andesitic rocks	57.3	59.4	Igneous
			Volcanic breccia, Tuff breccia	17.9	18.5	Sedimentary
			Porphyry	12.1	12.5	Igneous
			Rhyolitic rock	8.8	9.1	Igneous
			Rock tuff	3.9	4.1	Igneous

No.	Dam Catchment	Catchment Area (km ²)	Rock Types	Area (%)	Area (km ²)	Rock Classes
5	Kanayama	410.81	Pumice	29.7	120.0	Igneous
			Diabase rocks	9.9	40.0	Igneous
			Granitic rocks	2.0	8.0	Igneous
			Gabbro rocks	0.7	3.0	Igneous
			Hornfels	18.8	76.0	Metamorphic
			Gneiss	10.4	42.0	Metamorphic
			Slate	5.9	24.0	Metamorphic
			Quartzite Rock	0.7	3.0	Metamorphic
			Gravel, Sand, Clay	10.4	42.0	Sedimentary
			Clastic	8.2	33.0	Sedimentary
			Sandstone	1.2	5.0	Sedimentary
			Gravel, Sand	1.0	4.0	Sedimentary
			Sand, Gravel, Clay	0.7	3.0	Sedimentary
			Mudstone	0.3	1.0	Sedimentary
6	Nibutani	1155.45	Diabase rocks	13.8	158.9	Igneous
			Gabbro rocks	10.9	125.8	Igneous
			Granitic rocks	0.4	4.7	Igneous
			Slate	27.5	317.8	Metamorphic
			Serpentine rocks	10.6	121.9	Metamorphic
			Crystalline Schist	1.3	15.2	Metamorphic
			Gneiss	0.6	6.6	Metamorphic
			Mudstone	11.0	126.8	Sedimentary
			Alteration Sandstone, Mudstone	7.4	85.0	Sedimentary
			Gravel, Sand, Clay	5.9	67.9	Sedimentary
			Sandstone, Conglomerate	4.0	45.7	Sedimentary
			Gravel, Sand	3.3	38.1	Sedimentary
			Conglomerate	2.0	22.5	Sedimentary
			Sandstone	1.6	18.5	Sedimentary

No.	Dam Catchment	Catchment Area (km ²)	Rock Types	Area (%)	Area (km ²)	Rock Classes
7	Pirika	114.44	Granitic rocks	45.9	52.5	Igneous
			Rock tuff	18.9	21.7	Igneous
			Rhyolitic rock	0.2	0.2	Igneous
			Slate	19.2	21.9	Metamorphic
			Sandstone	7.8	8.9	Sedimentary
			Mudstone	4.9	5.6	Sedimentary
			Volcanic breccia, Tuff breccia	1.7	1.9	Sedimentary
			Gravel, Sand	1.2	1.4	Sedimentary
			Gravel, Sand, Clay	0.2	0.2	Sedimentary
8	Satsunaigawa	116.63	Granitic rocks	16.5	19.0	Igneous
			Gabbro rocks	7.8	9.0	Igneous
			Hornfels	56.5	65.0	Metamorphic
			Gneiss	8.7	10.0	Metamorphic
			Slate	1.7	2.0	Metamorphic
			Sandstone, Conglomerate	7.8	9.0	Sedimentary
			Gravel, Sand	0.9	1.0	Sedimentary
9	Taisetsu	289.26	Andesitic rocks	37.1	107.3	Igneous
			Pumice	7.3	21.1	Igneous
			Granitic rocks	2.7	7.8	Igneous
			Rock tuff	0.5	1.4	Igneous
			Slate	37.0	107.0	Metamorphic
			Hornfels	0.3	0.8	Metamorphic
			Sand, Gravel, Clay	6.1	17.6	Sedimentary
			Conglomerate	4.7	13.7	Sedimentary
			Gravel, Sand, Clay	2.1	6.1	Sedimentary
			Gravel, Sand	1.4	4.1	Sedimentary
			Clastic	0.7	2.1	Sedimentary
			Sandstone, Conglomerate	0.1	0.4	Sedimentary

No.	Dam Catchment	Catchment Area (km ²)	Rock Types	Area (%)	Area (km ²)	Rock Classes
10	Tokachi	598.24	Pumice	31.6	188.8	Igneous
			Andesitic rocks	22.9	137.0	Igneous
			Granitic rocks	1.1	6.3	Igneous
			Basaltic rocks	1.0	6.1	Igneous
			Rock tuff	0.0	0.2	Igneous
			Rhyolitic rock	0.0	0.2	Igneous
			Slate	38.9	232.9	Metamorphic
			Crystalline Schist	1.0	6.0	Metamorphic
			Gravel, Sand, Clay	1.9	11.6	Sedimentary
			Sandstone, Conglomerate	1.3	7.9	Sedimentary
			Gravel, Sand	0.2	1.0	Sedimentary
11	Gosho	635.17	Alteration of Rock	18.6	117.2	Igneous
			Andesitic rocks	14.7	93.0	Igneous
			Rhyolitic rock	14.1	89.2	Igneous
			Rock tuff	1.7	10.5	Igneous
			Granitic rocks	0.2	1.2	Igneous
			Volcanic Clastic Material	24.3	153.4	Sedimentary
			Tuff breccia and agglomerate	12.6	79.8	Sedimentary
			Mud, Silt, Gravel	10.4	65.4	Sedimentary
			Mudstone	2.8	17.6	Sedimentary
			Sandstone	0.7	4.3	Sedimentary
12	Sagurigawa	61.36	Mudstone	70.9	43.5	Sedimentary
			Volcanic Clastic Material	9.1	5.6	Sedimentary
			Mud, Silt, Gravel	7.6	4.6	Sedimentary
			Conglomerate, Sandstone, Mudstone	6.9	4.2	Sedimentary
			Pumice	5.6	3.4	Igneous
13	Ikari	271.2	Rhyolitic rock	36.1	99.0	Igneous
			Rock tuff	22.5	61.6	Igneous
			Granitic rocks	21.4	58.8	Igneous
			Andesitic rocks	0.3	0.9	Igneous
			Pumice	0.2	0.6	Igneous
			Quartzite Rock	0.2	0.4	Metamorphic
			Mudstone	18.7	51.3	Sedimentary
			Water bodies	0.6	1.7	

No.	Dam Catchment	Catchment Area (km ²)	Rock Types	Area (%)	Area (km ²)	Rock Classes
14	Kawaji	320.74	Pumice	1.4	4.4	Igneous
			Rock tuff	3.9	12.6	Igneous
			Rhyolitic rock	39.8	127.4	Igneous
			Andesitic rocks	18.9	60.4	Igneous
			Granitic rocks	13.0	41.6	Igneous
			Quartzite Rock	0.3	1.0	Metamorphic
			Mudstone	20.9	66.8	Sedimentary
			Schalstein	0.5	1.6	Sedimentary
			Limestone	0.5	1.6	Sedimentary
			Water bodies	0.9	3.0	
15	Kawamata	179.40	Rhyolitic rock	39.3	69.4	Igneous
			Granitic rocks	13.0	23.0	Igneous
			Andesitic rocks	8.6	15.1	Igneous
			Pumice	2.5	4.4	Igneous
			Rock tuff	0.1	0.2	Igneous
			Mudstone	34.0	60.0	Sedimentary
			Limestone	0.9	1.6	Sedimentary
			Water bodies	1.6	2.9	
16	Aimata	110.80	Rhyolitic rock	22.5	25.0	Igneous
			Andesitic rocks	22.4	24.9	Igneous
			Rock tuff	14.8	16.4	Igneous
			Granitic rocks	10.0	11.0	Igneous
			Mudstone	24.2	26.9	Sedimentary
			Gravel	2.8	3.1	Sedimentary
			Conglomerate	2.1	2.3	Sedimentary
			Water bodies	1.1	1.3	

No.	Dam Catchment	Catchment Area (km ²)	Rock Types	Area (%)	Area (km ²)	Rock Classes
17	Fujiwara	400.20	Rock tuff	1.0	3.8	Igneous
			Rhyolitic rock	5.8	22.7	Igneous
			Andesitic rocks	18.2	71.5	Igneous
			Granitic rocks	54.2	212.6	Igneous
			Gabbro rocks	0.4	1.7	Igneous
			Serpentine rocks	2.8	11.0	Metamorphic
			Gravel	0.4	1.5	Sedimentary
			Conglomerate	3.2	12.7	Sedimentary
			Sandstone	11.6	45.3	Sedimentary
			Mudstone	0.2	0.7	Sedimentary
			Loam	1.5	5.7	Sedimentary
			Water bodies	0.8	3.0	
18	Kusaki	263.85	Porphyry	27.6	72.9	Igneous
			Granitic rocks	27.6	72.8	Igneous
			Andesitic rocks	10.5	27.7	Igneous
			Rhyolitic rock	4.1	10.9	Igneous
			Hornfels	7.8	20.4	Metamorphic
			Quartzite Rock	0.1	0.4	Metamorphic
			Mudstone	22.2	58.5	Sedimentary
			Water bodies	0.1	0.2	
19	Naramata	95.4	Rhyolitic rock	0.1	0.1	Igneous
			Andesitic rocks	44.2	46.6	Igneous
			Granitic rocks	44.5	46.9	Igneous
			Serpentine rocks	7.5	7.9	Metamorphic
			Loam	3.8	4.0	Sedimentary

No.	Dam Catchment	Catchment Area (km ²)	Rock Types	Area (%)	Area (km ²)	Rock Classes
20	Shimokubo	323.65	Alteration of Rock	2.0	6.5	Igneous
			Quartzite Rock	18.8	60.7	Metamorphic
			Greenschist	6.1	19.8	Metamorphic
			Black Schist	0.2	0.5	Metamorphic
			Mudstone	55.3	178.9	Sedimentary
			Sandstone	15.7	50.9	Sedimentary
			Schalstein	1.7	5.6	Sedimentary
			Mud, Silt, Sand	0.2	0.6	Sedimentary
21	Sonohara	601.06	Andesitic rocks	37.9	228.0	Igneous
			Rhyolitic rock	31.2	187.3	Igneous
			Granitic rocks	8.5	51.0	Igneous
			Gabbro rocks	4.1	24.7	Igneous
			Porphyry	3.4	20.4	Igneous
			Serpentine rocks	0.9	5.5	Metamorphic
			Volcanic Clastic Material	3.6	21.9	Sedimentary
			Mudstone	3.2	19.1	Sedimentary
			Loam	3.0	18.2	Sedimentary
			Gravel	2.3	13.9	Sedimentary
			Sandstone	1.8	10.8	Sedimentary
			Water bodies	0.1	0.4	
22	Yagisawa	165.54	Granitic rocks	66.4	110.0	Igneous
			Rhyolitic rock	2.4	4.0	Igneous
			Gabbro rocks	1.0	1.7	Igneous
			Andesitic rocks	0.5	0.9	Igneous
			Sandstone	27.4	45.3	Sedimentary
			Mudstone	0.4	0.7	Sedimentary
			Water bodies	1.8	3.0	

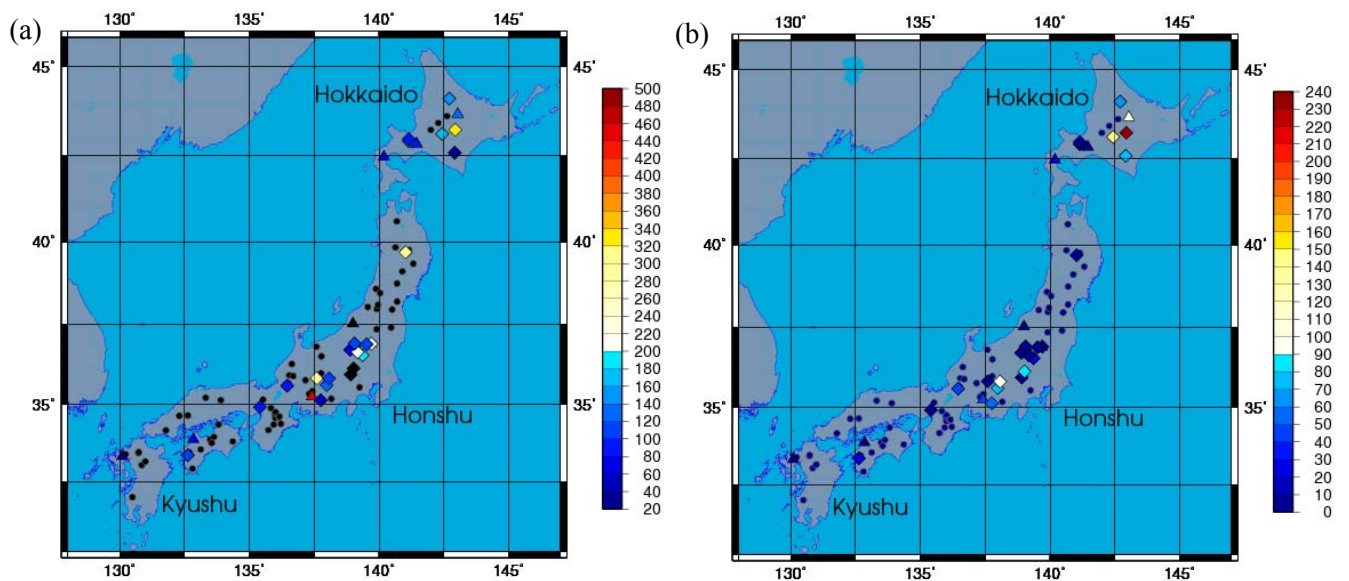
No.	Dam Catchment	Catchment Area (km ²)	Rock Types	Area (%)	Area (km ²)	Rock Classes
23	Futase	170.58	Granitic rocks	4.9	8.3	Igneous
			Alteration of Rock	0.0	0.1	Igneous
			Hornfels	0.7	1.2	Metamorphic
			Quartzite Rock	0.5	0.9	Metamorphic
			Mudstone	61.5	104.9	Sedimentary
			Sandstone	30.3	51.7	Sedimentary
			Limestone	1.8	3.1	Sedimentary
			Clastic	0.2	0.3	Sedimentary
			Schalstein	0.1	0.1	Sedimentary
24	Koshibu	289.57	Alteration of Rock	26.7	77.3	Igneous
			Granitic rocks	18.7	54.0	Igneous
			Greenschist	8.3	24.1	Metamorphic
			Serpentine rocks	5.7	16.4	Metamorphic
			Black Schist	4.4	12.6	Metamorphic
			Quartzite Rock	4.1	11.7	Metamorphic
			Hornfels	3.6	10.4	Metamorphic
			Mudstone	20.7	59.8	Sedimentary
			Limestone	4.2	12.0	Sedimentary
			Mud, Silt, Gravel	0.5	1.5	Sedimentary
			Gravel, Sand	0.0	0.0	Sedimentary
			Rock crushing	3.3	9.5	
25	Makio	307.79	Rhyolitic rock	50.8	156.4	Igneous
			Andesitic rocks	33.1	101.9	Igneous
			Alteration of Rock	10.2	31.3	Igneous
			Basaltic rocks	1.1	3.5	Igneous
			Granitic rocks	0.9	2.8	Igneous
			Mud, Silt, Gravel	3.4	10.3	Sedimentary
			Conglomerate	0.5	1.6	Sedimentary

No.	Dam Catchment	Catchment Area (km ²)	Rock Types	Area (%)	Area (km ²)	Rock Classes
26	Miwa	311.03	Alteration of Rock	30.6	95.1	Igneous
			Granitic rocks	3.4	10.5	Igneous
			Gabbro rocks	0.3	0.8	Igneous
			Black Schist	15.4	47.7	Metamorphic
			Greenschist	7.1	22.0	Metamorphic
			Hornfels	5.9	18.3	Metamorphic
			Crystalline Schist	1.5	4.7	Metamorphic
			Serpentine rocks	0.3	0.9	Metamorphic
			Mudstone	24.5	76.1	Sedimentary
			Limestone	9.1	28.3	Sedimentary
			Mud, Silt, Gravel	0.6	1.8	Sedimentary
			Gravel, Sand	0.0	0.1	Sedimentary
			Rock crushing	1.5	4.7	
27	Maruyama	2409.00	Alteration of Rock	21.1	510.8	Igneous
			Rhyolitic rock	26.2	633.6	Igneous
			Andesitic rocks	7.2	173.0	Igneous
			Basaltic rocks	0.8	18.9	Igneous
			Porphyry	1.2	27.8	Igneous
			Granitic rocks	27.4	662.3	Igneous
			Quartzite Rock	0.9	21.3	Metamorphic
			Hornfels	0.9	22.5	Metamorphic
			Gravel	3.5	83.9	Sedimentary
			Gravel, Sand	0.8	18.8	Sedimentary
			Mud, Silt, Gravel	3.8	91.0	Sedimentary
			Gravel, Sand, Clay	2.7	65.0	Sedimentary
			Conglomerate	0.1	1.6	Sedimentary
			Sandstone	3.4	81.5	Sedimentary
			Mudstone	0.1	3.4	Sedimentary
			Tuff breccia and agglomerate	0.2	4.7	Sedimentary
			Water bodies	0.1	1.6	

No.	Dam Catchment	Catchment Area (km ²)	Rock Types	Area (%)	Area (km ²)	Rock Classes
28	Yokoyama	470.71	Granitic rocks	13.2	62.1	Igneous
			Alteration of Rock	5.1	24.2	Igneous
			Quartzite Rock	9.0	42.4	Metamorphic
			Sandstone	30.9	145.5	Sedimentary
			Mudstone	21.4	100.6	Sedimentary
			Schalstein	18.8	88.6	Sedimentary
			Limestone	0.8	3.8	Sedimentary
			Conglomerate	0.7	3.2	Sedimentary
			Tuff breccia and agglomerate	0.1	0.4	Sedimentary
29	Sintoyone	111.44	Granitic rocks	37.7	42.0	Igneous
			Andesitic rocks	4.2	4.7	Igneous
			Alteration of Rock	4.1	4.5	Igneous
			Rock tuff	2.6	2.9	Igneous
			Crystalline Schist	49.0	54.6	Metamorphic
			Hornfels	1.9	2.1	Metamorphic
			Gravel, Sand, Clay	0.6	0.6	Sedimentary
30	Yahagi	504.62	Granitic rocks	87.1	439.7	Igneous
			Andesitic rocks	0.8	4.1	Igneous
			Hornfels	7.9	40.0	Metamorphic
			Crystalline Schist	0.1	0.6	Metamorphic
			Gravel, Sand, Clay	2.8	14.2	Sedimentary
			Sandstone	0.9	4.5	Sedimentary
			Mud, Silt, Gravel	0.3	1.6	Sedimentary

No.	Dam Catchment	Catchment Area (km ²)	Rock Types	Area (%)	Area (km ²)	Rock Classes
31	Hitokura	115.10	Granitic rocks	39.8	45.2	Igneous
			Alteration of Rock	26.3	29.9	Igneous
			Rhyolitic rock	5.8	6.6	Igneous
			Sand	11.2	12.7	Sedimentary
			Sandstone	8.4	9.5	Sedimentary
			Mudstone	7.1	8.0	Sedimentary
			Mud, Silt, Gravel	1.4	1.6	Sedimentary
32	Ishitegawa	72.60	Granitic rocks	87.5	63.8	Igneous
			Alteration of Rock	6.1	4.5	Igneous
			Hornfels	6.4	4.7	Metamorphic
33	Nomura	168.00	Alteration of Rock	66.1	112.8	Igneous
			Quartzite Rock	12.8	21.9	Metamorphic
			Mud, Silt, Gravel	21.1	35.9	Sedimentary
34	Kyuuragi	33.70	Granitic rocks	92.8	31.8	Igneous
			Greenschist	7.2	2.5	Metamorphic

After classifying those rock types into rock classes, the geographical distribution of each rock class is shown in **Figure 5.2**, and the percentage area of rock classes for each catchment is presented in **Table 5.3**.



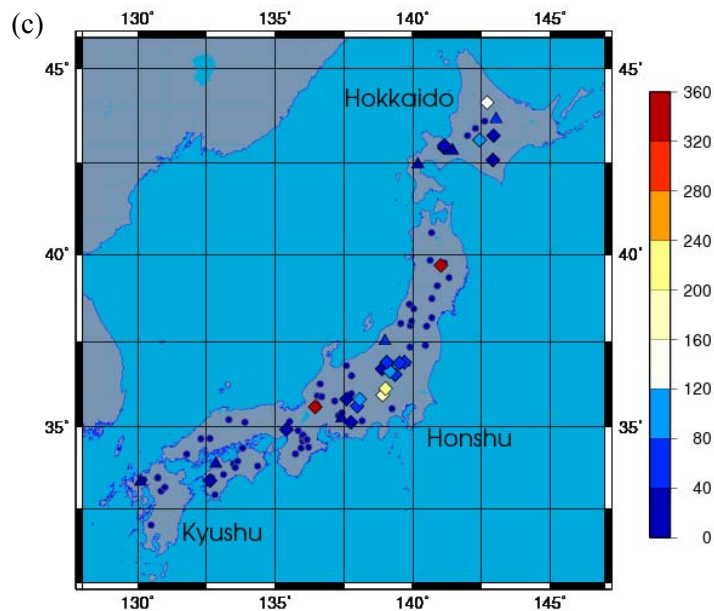


Figure 5.2. Geographic distribution of rock class area in km² for (a) igneous rock class, (b) metamorphic rock class, (c) sedimentary rock class.

Table 5.3. Percentage area of rock classes for each catchment.

No	Dam Catchment	Catchment Area (km ²)	Igneous Rock Area (%)	Metamorphic Rock Area (%)	Sedimentary Rock Area (%)
1	Houheikyou	136.1	91.4	0.0	8.6
2	Jyouzankei	103.6	82.1	0.0	17.9
3	Kanayama	410.8	42.3	35.9	21.8
4	Satsunaigawa	116.6	24.4	67.0	8.7
5	Iwaonai	341.6	45.0	19.9	35.2
6	Tokachi	598.2	56.6	40.0	3.4
7	Pirika	114.4	65.0	19.2	15.8
8	Taisetsu	289.3	47.6	37.3	15.2
9	Nibutani	1155.5	25.0	40.0	35.0
10	Izarigawa	113.3	79.8	0.0	20.2
11	Gosho	635.2	49.3	0.0	50.8
12	Sagurigawa	61.4	5.6	0.0	94.4
13	Ikari	271.2	80.5	0.2	18.7
14	Kawamata	179.4	63.5	0.0	34.9
15	Aimata	110.8	69.7	0.0	29.2
16	Kusaki	263.9	69.9	7.9	22.2
17	Sonohara	601.1	85.1	0.9	14.0
18	Yagisawa	165.5	70.4	0.0	27.8
19	Futase	170.6	4.9	1.2	93.9
20	Shimokubo	323.7	2.0	25.1	73.0
21	Makio	307.8	96.1	0.0	3.9
22	Miwa	311.0	34.2	30.1	34.2
23	Yokoyama	470.7	18.3	9.0	72.7
24	Sintoyone	111.4	48.5	50.9	0.6
25	Yahagi	504.6	87.9	8.0	4.0
26	Koshibu	289.6	45.4	26.0	25.3
27	Hitokura	115.1	71.9	0.0	28.1
28	Ishitegawa	72.6	93.6	6.4	0.0
29	Nomura	168.0	66.1	12.8	21.1
30	Kyuuragi	33.7	92.8	7.2	0.0
31	Kawaji	320.7	76.9	0.3	21.9
32	Fujiwara	400.2	79.6	2.8	16.8
33	Naramata	95.4	88.8	7.5	3.8
34	Maruyama	2409.0	83.7	1.8	14.4

After calculating the percentage area of rock classes for each catchment, the scatter diagram between fraction of catchment area, and fraction of minimum total rainfall required to generate direct runoff for catchments having a constant-stage tanh-type curve is presented in **Figure 5.3**.

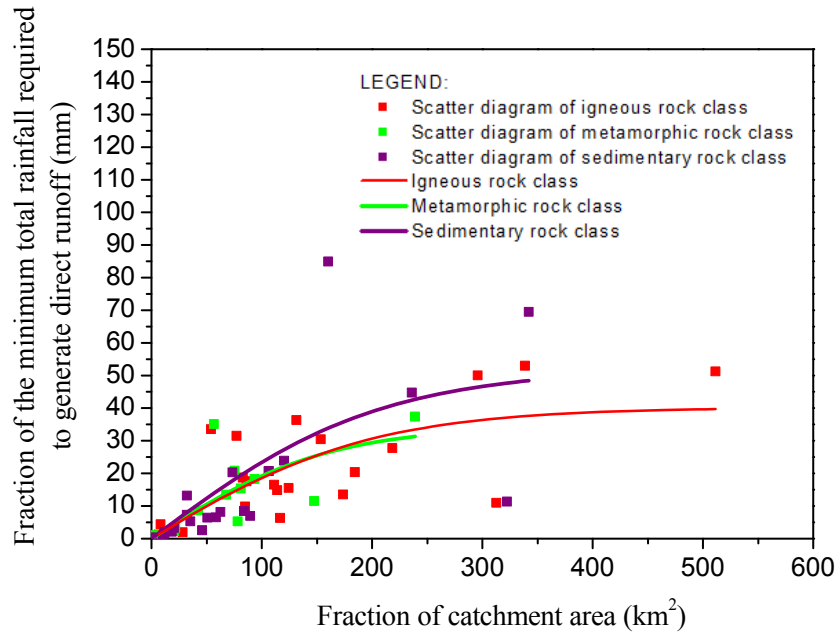


Figure 5.3. Scatter diagram between fraction of catchment area (km^2) and fraction of the minimum total rainfall required to generate direct runoff (mm) for catchments having a constant-stage tanh-type curve. Red denotes igneous rocks, green denotes metamorphic rocks, and purple denotes sedimentary rocks.

Sedimentary rocks tend to have high primary porosity and very high hydraulic conductivity compared to igneous and metamorphic rocks (Fetter, 2000; Weight, 2008). Thus, catchments with a larger area of sedimentary rocks are permeable, and rainwater can easily infiltrate. This explains why catchment areas with a larger fraction of sedimentary rocks tend to require higher minimum total rainfall to generate direct runoff, followed by those areas that are dominated by igneous rocks and metamorphic rocks (see **Figure 5.3**). However, the relationship between the fraction of catchment area of a given rock class and the minimum total rainfall to generate direct runoff is non-linear, that is possibly due to different hydraulic conductivity of each rock type (Weight, 2008), rainfall spatial distribution, and initial conditions of soil moisture (Supraba and Yamada, 2014).

Figure 5.3 shows that catchments with a larger area of igneous rocks tend to require the next highest minimum total rainfall to generate direct runoff, after those with a larger area of sedimentary rocks. Igneous rocks are an important water source in some regions (Weight, 2008). Igneous rocks such as basalt, andesite, and rhyolite, have a high capacity for water transmission to transfer the rain falling on the ground to the underground due to their high permeability, and for water storage (Weight, 2008). This is why the majority of catchments, where the minimum total rainfall required to generate direct runoff was > 30 mm, occurred in catchments covered with igneous rocks.

5.3 SUMMARY

The new findings found on this chapter can be summarized as follows:

- The minimum total rainfall required to generate direct runoff, and total rainfall loss under saturation condition are important parameters to estimate the potential catchment storage. This study proposed the threshold to quantify the minimum total rainfall required to generate direct runoff as at least 95% of total rainfall become total rainfall loss.
- There are 36 rock types in Japan, and igneous rock class is the most dominant rock class in Japan.
- Variation in minimum total rainfall required to generate direct runoff, and in total rainfall loss under saturation condition can be explained by rock class, drainage density, elongation ratio, and catchment width. Catchments covered up by a bigger fraction of rocks belongs to sedimentary rock class tend to have bigger value of minimum total rainfall required to generate direct runoff and bigger value of total rainfall loss under saturation condition. Catchments having higher drainage density, elongation ratio, and catchment width tends to have higher direct runoff, so those catchments will have smaller value of minimum total rainfall required to generate direct runoff, and smaller value of total rainfall loss under saturation condition.

Chapter 6. UNCERTAINTY OF PEAK RUNOFF

This chapter is written based on a published paper:

Intan Supraba and Tomohito J. Yamada, “Uncertainty of Peak Runoff Height Associated With Uncertainty of Water Holding Capacity and Rainfall Pattern”, *Journal of Global Environmental Engineering*, 2015.

The content of this chapter has been presented in the following conference:

Intan Supraba and Tomohito J. Yamada, “Uncertainty of Peak Runoff Height Associated With Uncertainties of Water Holding Capacity and Rainfall Pattern”, in Japan Geoscience Union (JpGU) Meeting 2015, 24th-28th May 2015, Makuhari Messe, Japan.

6.1 HISTORY OF STOCHASTIC DIFFERENTIAL EQUATION

(Robert Brown, 1827) observed the random movement of particles suspended in fluid but he could not explained the mechanisms that caused this motion. (Albert Einstein, 1905) interpreted that the irregular (random) motion observed by Brown was a result of the particle being moved by individual water molecules due to the molecular kinetic theory of heat, and the motion is named Brownian motion. Thus Brownian motion is described as the random motion of particles suspended in a fluid because of collision with the molecules in the fluid. The collision cause the transfer of the particle momentum to the molecules of the fluid, thus the velocity of the particle decreases to zero. The equation of motion for the particle is expressed as follows:

$$m\dot{v} + \alpha v = 0 \quad (58)$$

The differential equation in Eq. (58) is a deterministic equation because the velocity $v(t)$ at time t is completely determined by its initial value. The deterministic equation is valid only if the mass of the particle is large so its velocity due to thermal fluctuations can be neglected.

However, Eq. (58) needs to be modified so it leads to the correct thermal energy by adding a fluctuation force $F_f(t)$. This force $F_f(t)$ is a stochastic or random force. The total force of the molecules acting on the small particle consists of a continuous damping force $F_c(t)$ and a fluctuation force $F_f(t)$:

$$F(t) = F_c(t) + F_f(t) = -\alpha v(t) + F_f(t) \quad (59)$$

The equation of motion can be obtained by inserting $F_f(t)$ into Eq. (58):

$$m\dot{v} + \alpha v = F_f(t) \quad (60)$$

After dividing Eq. (60) by mass:

$$\dot{v} + \frac{\alpha}{m} v = \frac{F_f(t)}{m} \quad (61)$$

Eq. (61) can be written as:

$$\dot{v} + \gamma v = \Gamma(t) \quad (62)$$

where:

$$\gamma = \frac{\alpha}{m} \text{ and } \frac{F_f(t)}{m} = \Gamma(t) \quad (63)$$

$\Gamma(t)$ is the fluctuating force per unit mass that is called the Langevin force. Eq. (63) contains the stochastic force so it is called a stochastic differential equation. The Langevin force has some properties as follows:

- The average of Langevin force over the ensemble should be zero because the equation of motion of the average velocity $\overline{v(t)}$ should be given by Eq. (58):

$$\overline{\Gamma(t)} = 0 \quad (64)$$

- The average value of multiplication of two Langevin force at different times is zero for time differences $t' - t$ which are larger than the duration time τ_0 of a collision because the collisions of different molecules of the fluid with the small particle are approximately independent:

$$\overline{\Gamma(t)\Gamma(t')} = 0 \text{ for } |t - t'| \geq \tau_0 \quad (65)$$

- The duration time τ_0 of a collision is much smaller than the relaxation time $\tau = 1/\gamma$ of the velocity of the small particle. Thus, by taking the limit $\tau_0 \rightarrow 0$:

$$\overline{\Gamma(t)\Gamma(t')} = q\delta(t - t') \quad (66)$$

The mean energy of the particle based on the equipartition law:

$$\frac{1}{2}m\overline{v^2} = \frac{1}{2}kT \quad (67)$$

According to the **Eq. (67)**, the average energy of the small particle cannot be finite so the δ function appears in the **Eq. (66)**. And the noise strength q of the Langevin force is expressed as follows:

$$q = 2\gamma kT/m \quad (68)$$

It is usually assumed that the $\Gamma(t)$ has a Gaussian distribution with δ correlation. Then the diffusion constant can be calculated by integrating **Eq. (62)**, and by using **Eq. (64), (66), and (68)**. A noise force with the δ correlation is called white noise, whereas the noise force without δ correlation is called colored noise.

From **Eq. (62)**, $\Gamma(t)$ varies from system to system (stochastic quantity), and the velocity v also varies from system to system. The velocity v is a continuous variable, and the probability density or probability distribution $W(v)$ multiplied by the interval dv is the probability of finding the particle

in the interval $(v, v + dv)$. The distribution function depends on time t and the initial distribution. Thus, the equation of motion for the distribution function $W(v, t)$ can be expressed as follows:

$$\frac{\partial W}{\partial t} = \gamma \frac{\partial (vW)}{\partial v} + \gamma \frac{kT}{m} \frac{\partial^2 W}{\partial v^2} \quad (69)$$

Eq. (69) is one of the simplest Fokker-Planck equations. And the general Fokker-Planck equation for one variable x is expressed as follows:

$$\frac{\partial W}{\partial t} = \left[-\frac{\partial}{\partial x} D^{(1)}(x) + \frac{\partial^2}{\partial x^2} D^{(2)}(x) \right] W \quad (70)$$

where: $D^{(1)}(x)$ is the drift coefficient, and $D^{(2)}(x)$ is the diffusion coefficient.

6.2 RELATIONSHIP BETWEEN ITO STOCHASTIC DIFFERENTIAL EQUATION AND FOKKER-PLANCK EQUATION

The well-known stochastic differential equation is Ito Stochastic Differential Equation is expressed as follows (Ito, 1950):

$$dx(t) = \mu(x(t), t)dt + \sigma(x(t), t)dW(t) \quad (71)$$

where x is the probabilistic or random variable, $dx(t)$ is the increment of probabilistic variable for each time step, $\mu(x(t), t)dt$ is the drift or deterministic term; $\mu(x(t))$ is the drift coefficient; $\sigma(x(t), t)dW(t)$ is the martingale term; $\sigma(x(t))$ is the volatile/ random coefficient; and $dW(t)$ or Wiener process that is known as standard Brownian motion is the increment of a continuous time stochastic process, and the increments for non-overlapping time intervals are independent.

The Fokker-Planck equation can be obtained by expanding **Eq. (71)** based on the following procedure. The standard normal distribution has probability density:

$$f(x) = \frac{1}{\sqrt{2\pi}} e^{-x^2/2} \quad (72)$$

If a random variable x is given and its distribution admits a probability density function f , then the expected value of x can be calculated as:

$$E(x) = \int_{-\infty}^{\infty} x f(x) dx \quad (73)$$

The Taylor series of a real function $f(x)$ that is infinitely differentiable at a real number x_0 is the power series:

$$\begin{aligned} f(x) = f(x_0) &+ \left(\frac{df}{dx}\right)_{x=x_0} (x - x_0) + \frac{1}{2!} \left(\frac{d^2f}{dx^2}\right)_{x=x_0} (x - x_0)^2 \\ &+ \frac{1}{3!} \left(\frac{d^3f}{dx^3}\right)_{x=x_0} (x - x_0)^3 + \dots + \frac{1}{k!} \left(\frac{d^kf}{dx^k}\right)_{x=x_0} (x - x_0)^k \end{aligned} \quad (74)$$

After rearranging Eq. (74):

$$\begin{aligned} f(x) - f(x_0) &= \left(\frac{df}{dx}\right)_{x=x_0} (x - x_0) + \frac{1}{2!} \left(\frac{d^2f}{dx^2}\right)_{x=x_0} (x - x_0)^2 \\ &+ \frac{1}{3!} \left(\frac{d^3f}{dx^3}\right)_{x=x_0} (x - x_0)^3 + \dots + \frac{1}{k!} \left(\frac{d^kf}{dx^k}\right)_{x=x_0} (x - x_0)^k \end{aligned} \quad (75)$$

From Eq. (75):

$$df(x) = f'(x) dx + \frac{1}{2} f''(x) (dx)^2 + \dots + \frac{1}{k!} f^k(x) (dx)^k \quad (76)$$

where $dx = (x - x_0)$.

$$\text{For } k \geq 3, (dx)^k = 0 \quad (77)$$

After substituting Eq. (77) into Eq. (76):

$$df(x) = f'(x) dx + \frac{1}{2} f''(x) (dx)^2 \quad (78)$$

The drift and random coefficients in Eq. (71) are assumed to be represented by the following variables:

$$\mu(x(t)) = y \text{ and } \sigma(x(t)) = z \quad (79)$$

After substituting Eq. (79) into Eq. (71):

$$dx(t) = y(t)dt + z(t)dW(t) \quad (80)$$

By taking the square of Eq. (80):

$$(dx(t))^2 = (y(t)dt + z(t)dW(t))^2 \quad (81)$$

Expand the right side of Eq. (81):

$$(y(t)dt + z(t)dW(t))^2 = (y(t))^2(dt)^2 + 2y(t)dt z(t)dW(t) + (z(t))^2(dW(t))^2 \quad (82)$$

From Eq. (82):

$$(y(t)dt + z(t)dW(t))^2 = (z(t))^2(dW(t))^2 \quad (83)$$

From Eq. (81) and Eq. (83):

$$(dx(t))^2 = (z(t))^2(dW(t))^2 \quad (84)$$

$$dW(t)^2 = dt \quad (85)$$

After substituting Eq. (85) into Eq. (84):

$$(dx(t))^2 = (z(t))^2 dt \quad (86)$$

After substituting Eq. (79) into Eq. (86):

$$(dx(t))^2 = (\sigma(x(t)), t)^2 dt \quad (87)$$

Then after substituting Eq. (71) and Eq. (87) into Eq. (78), and convert it into partial differential equation:

$$\begin{aligned} \partial f(x(t), t) &= f'(x(t))[\mu(x(t), t)\partial t + \sigma(x(t), t)dW(t)] \\ &+ \frac{1}{2}f''(x(t))(\sigma(x(t)), t)^2 \partial t \end{aligned} \quad (88)$$

After rearranging Eq. (88):

$$\begin{aligned}\partial f(x(t), t) = & f'(x(t))\mu(x(t), t)\partial t + f'(x(t))\sigma(x(t), t)dW(t) \\ & + \frac{1}{2}f''(x(t))(\sigma(x(t)), t)^2\partial t\end{aligned}\quad (89)$$

Then after rearranging Eq. (89):

$$\begin{aligned}\partial f(x(t), t) = & \left[f'(x(t))\mu(x(t), t) \right. \\ & \left. + \frac{1}{2}f''(x(t))(\sigma(x(t)), t)^2 \right] \partial t + f'(x(t))\sigma(x(t), t)dW(t)\end{aligned}\quad (90)$$

By renaming f in Eq. (90) into q :

$$\begin{aligned}\partial q(x(t), t) = & \left[q'(x(t))\mu(x(t), t) \right. \\ & \left. + \frac{1}{2}q''(x(t))(\sigma(x(t)), t)^2 \right] \partial t + q'(x(t))\sigma(x(t), t)dW(t)\end{aligned}\quad (91)$$

From Eq. (72), if $x = q(x(t))$ and the probability density function $f(x) = p(x(t))$, then substitute it into Eq. (73):

$$E(q(x(t))) = \int_{-\infty}^{\infty} q(x(t))p(x(t))dx(t)\quad (92)$$

Converting Eq. (92) into partial differential equation:

$$E(q(x(t), t)) = \int_{-\infty}^{\infty} q(x(t), t)p(x(t), t)\partial(x(t), t)\quad (93)$$

$$\frac{\partial}{\partial t} E[q(x(t), t)] = E\left[\frac{\partial q(x(t), t)}{\partial t}\right]\quad (94)$$

$$E[dW(t)] = 0\quad (95)$$

After substituting Eq. (91) into Eq. (94):

$$\begin{aligned} & \frac{\partial}{\partial t} E[q(x(t), t)] \\ &= E \left[\frac{\left[q'(x(t))\mu(x(t), t) + \frac{1}{2} q''(x(t))(\sigma(x(t)), t)^2 \right] \partial t + q'(x(t))\sigma(x(t), t)dW(t)}{\partial t} \right] \end{aligned} \quad (96)$$

After substituting Eq. (95) into Eq. (96):

$$\frac{\partial}{\partial t} E[q(x(t), t)] = E \left[\frac{\left[q'(x(t))\mu(x(t), t) + \frac{1}{2} q''(x(t))(\sigma(x(t)), t)^2 \right] \partial t}{\partial t} \right] \quad (97)$$

After rearranging Eq. (97):

$$\frac{\partial}{\partial t} E[q(x(t), t)] = E \left[q'(x(t))\mu(x(t), t) + \frac{1}{2} q''(x(t))(\sigma(x(t)), t)^2 \right] \quad (98)$$

After substituting Eq. (93) into the left side of Eq. (98):

$$\frac{\partial}{\partial t} E[q(x(t), t)] = \frac{\partial}{\partial t} \int_{-\infty}^{\infty} q(x(t), t) p(x(t), t) \partial(x(t), t) \quad (99)$$

After rearranging Eq. (99):

$$\frac{\partial}{\partial t} E[q(x(t), t)] = \int_{-\infty}^{\infty} q(x(t), t) \frac{\partial p(x(t), t)}{\partial t} \partial(x(t), t) \quad (100)$$

From Eq. (100):

$$E[q(x(t), t)] = q(x(t), t) \int_{-\infty}^{\infty} p(x(t), t) \partial(x(t), t) \quad (101)$$

After rearranging Eq. (101):

$$E = \int_{-\infty}^{\infty} p(x(t), t) \partial(x(t), t) \quad (102)$$

After substituting Eq. (102) into the right side of Eq. (98):

$$\int_{-\infty}^{\infty} p(x(t), t) \partial(x(t), t) \left[q'(x(t)) \mu(x(t), t) + \frac{1}{2} q''(x(t)) (\sigma(x(t)), t)^2 \right] \quad (103)$$

After rearranging Eq. (103):

$$\int_{-\infty}^{\infty} \left[q'(x(t)) \mu(x(t), t) + \frac{1}{2} q''(x(t)) (\sigma(x(t)), t)^2 \right] p(x(t), t) \partial(x(t), t) \quad (104)$$

From Eq. (98), left side = right side $\rightarrow$ Eq. (100) = Eq. (104):

$$\begin{aligned} & \int_{-\infty}^{\infty} q(x(t), t) \frac{\partial p(x(t), t)}{\partial t} \partial(x(t), t) \\ &= \int_{-\infty}^{\infty} \left[q'(x(t)) \mu(x(t), t) \right. \\ & \quad \left. + \frac{1}{2} q''(x(t)) (\sigma(x(t)), t)^2 \right] p(x(t), t) \partial(x(t), t) \end{aligned} \quad (105)$$

$$q'(x(t)) = \frac{\partial q(x(t), t)}{\partial(x(t))}; \quad q''(x(t)) = \frac{\partial^2 q(x(t), t)}{\partial x(t)^2} \quad (106)$$

After substituting Eq. (106) into Eq. (105):

$$\begin{aligned}
& \int_{-\infty}^{\infty} q(x(t), t) \frac{\partial p(x(t), t)}{\partial t} \partial(x(t), t) \\
&= \int_{-\infty}^{\infty} \left[\frac{\partial q(x(t), t)}{\partial(x(t))} \mu(x(t), t) \right. \\
&\quad \left. + \frac{1}{2} \frac{\partial^2 q(x(t), t)}{\partial x(t)^2} (\sigma(x(t)), t)^2 \right] p(x(t), t) \partial(x(t), t)
\end{aligned} \tag{107}$$

After rearranging Eq. (107):

$$\begin{aligned}
& \frac{\partial p(x(t), t)}{\partial t} \int_{-\infty}^{\infty} q(x(t), t) \partial(x(t), t) \\
&= \int_{-\infty}^{\infty} \left[\frac{\partial q(x(t), t)}{\partial(x(t))} \mu(x(t), t) p(x(t), t) \right. \\
&\quad \left. + \frac{1}{2} \frac{\partial^2 q(x(t), t)}{\partial x(t)^2} (\sigma(x(t)), t)^2 p(x(t), t) \right] \partial(x(t), t)
\end{aligned} \tag{108}$$

From Eq. (108):

$$\begin{aligned}
& \frac{\partial p(x(t), t)}{\partial t} \\
&= \frac{\int_{-\infty}^{\infty} \left[\frac{\partial q(x(t), t)}{\partial(x(t))} \mu(x(t), t) p(x(t), t) + \frac{1}{2} \frac{\partial^2 q(x(t), t)}{\partial x(t)^2} (\sigma(x(t)), t)^2 p(x(t), t) \right] \partial(x(t), t)}{\int_{-\infty}^{\infty} q(x(t), t) \partial(x(t), t)}
\end{aligned} \tag{109}$$

From Eq. (109):

$$\begin{aligned}
& \frac{\partial p(x(t), t)}{\partial t} = \frac{\int_{-\infty}^{\infty} \frac{\partial q(x(t), t)}{\partial(x(t))} \mu(x(t), t) p(x(t), t) \partial(x(t), t)}{\int_{-\infty}^{\infty} q(x(t), t) \partial(x(t), t)} \\
&\quad + \frac{\int_{-\infty}^{\infty} \frac{1}{2} \frac{\partial^2 q(x(t), t)}{\partial x(t)^2} (\sigma(x(t)), t)^2 p(x(t), t) \partial(x(t), t)}{\int_{-\infty}^{\infty} q(x(t), t) \partial(x(t), t)}
\end{aligned} \tag{110}$$

From Eq. (110):

$$\frac{\partial p(x(t), t)}{\partial t} = -\frac{\partial \mu(x(t), t)p(x(t), t)}{\partial (x(t))} + \frac{1}{2} \frac{\partial^2 (\sigma(x(t)), t)^2 p(x(t), t)}{\partial x(t)^2} \quad (111)$$

Eq. (111) is the Fokker-Planck equation where $p(x, t)$ is the probability density function in phase space.

6.3 PROPOSED METHODS TO QUANTIFY UNCERTAINTY OF PEAK RUNOFF HEIGHT

Section 3.2 has discussed about the standard deviation of parameter a that reflects the uncertainty of water holding capacity. Thus, this uncertainty of water holding capacity affects the uncertainty of effective rainfall that results in the uncertainty of peak runoff height. A recent study identified the uncertainty of peak runoff height due to uncertainty of rainfall distribution based on stochastic differential equation method. Thus, the main purpose of this study is to investigate the uncertainty of peak runoff height by considering uncertainty of effective rainfall through the water holding capacity in addition to uncertainty of rainfall distribution by proposing two methods namely ensemble method and stochastic differential equation (sde) method.

The uncertainty of peak runoff height is quantified by considering two different types of uncertainty i.e. uncertainty associated with rainfall distribution, and uncertainty associated with water holding capacity. Two methods are proposed to quantify the uncertainty of peak runoff height i.e. ensemble method and sde method.

6.3.1 ENSEMBLE METHOD

Total rainfall loss against target total rainfall behaves Gaussian. Thus, 103 cases of runoff parameters are estimated by considering that parameter a behaves Gaussian (normal distribution). Then, each of cases is used to simulate runoff height by using **Eq. (31)**. The uncertainty of peak runoff height is quantified by subtracting the maximum and minimum peak runoff height among 103 cases.

6.3.2 STOCHASTIC DIFFERENTIAL EQUATION METHOD

In this method, the uncertainty of peak runoff height is quantified by modifying the effective rainfall intensity as follows:

$$r_e(t) = \bar{r}_e(t) + r_e'(t) \quad (112)$$

where $\bar{r}_e(t)$ is the effective rainfall intensity based on the deterministic value of runoff parameter a (mm/h), and $r_e'(t)$ is the uncertainty that contributes to the uncertainty of peak runoff height (mm/h). Then the uncertainty of peak runoff height can be obtained by substituting Eq. (112) into Eq. (31):

$$\frac{dq_*(t)}{dt} = a_0 q_*(t)^\beta (\bar{r}_e(t) + r_e'(t) - q_*(t)) \quad (113)$$

After rearranging and multiplying Eq. (113) by dt :

$$dq_*(t) = a_0 q_*(t)^\beta (\bar{r}_e(t) - q_*(t))dt + a_0 q_*(t)^\beta r_e'(t)dt \quad (114)$$

The random term in Eq. (71) i.e. $\sigma(x(t), t)dW(t)$ is analogous to the term $r_e'(t)dt$ in Eq. (114). Thus, after replacing $r_e'(t)dt$ with $\sigma(x(t), t)dW(t)$:

$$dq_*(t) = a_0 q_*(t)^\beta (\bar{r}_e(t) - q_*(t))dt + a_0 q_*(t)^\beta \sigma(x(t), t)dW(t) \quad (115)$$

Eq. (115) can be written as:

$$dq_*(t) = \mu(q_*(t))dt + f(q_*(t))dW(t) \quad (116)$$

From Eq. (115) and Eq. (116):

$$\mu(q_*(t)) \text{ is } a_0 q_*(t)^\beta (\bar{r}_e(t) - q_*(t)) \text{ and } f(q_*(t)) \text{ is } a_0 q_*(t)^\beta \sigma(x(t), t) \quad (117)$$

From Eq. (71) and Eq. (116):

$$x(t) \approx q_*(t) \text{ and } \sigma(x(t), t) \approx f(q_*(t)) \quad (118)$$

After substituting Eq. (118) into Eq. (111):

$$\frac{\partial p(q_*(t), t)}{\partial t} = -\frac{\partial \mu(q_*(t)) p(q_*(t), t)}{\partial q_*(t)} + \frac{1}{2} \frac{\partial^2 (f(q_*(t)))^2 p(q_*(t), t)}{\partial q_*(t)^2} \quad (119)$$

Then after substituting Eq. (117) into Eq. (119):

$$\begin{aligned} \frac{\partial p(q_*(t), t)}{\partial t} = & - \frac{\partial a_0 q_*(t)^\beta (\bar{r}_e(t) - q_*(t)) p(q_*(t), t)}{\partial q_*(t)} \\ & + \frac{1}{2} \frac{\partial^2 \left(a_0 q_*(t)^\beta \sigma(x(t), t) \right)^2 p(q_*(t), t)}{\partial q_*(t)^2} \end{aligned} \quad (120)$$

Replacing $\sigma(x(t), t)$ with $r_e'(t)$:

$$\begin{aligned} \frac{\partial p(q_*(t), t)}{\partial t} = & - \frac{\partial a_0 q_*(t)^\beta (\bar{r}_e(t) - q_*(t)) p(q_*(t), t)}{\partial q_*(t)} \\ & + \frac{1}{2} \frac{\partial^2 \left(a_0 q_*(t)^\beta r_e'(t) \right)^2 p(q_*(t), t)}{\partial q_*(t)^2} \end{aligned} \quad (121)$$

Eq. (121) is the equation of probability of runoff height. This equation is analogous to the advection-diffusion equation where the random term (uncertainty) in Eq. (121) is similar to the diffusion term in the advection-diffusion equation. Then Eq. (121) is solved numerically by giving the initial condition based on Runge-Kutta methods, then choose the solution at time to peak to get the probability of peak runoff height.

In this study the uncertainty that contributes to the uncertainty of peak runoff height, $r_e'(t)$, is considered for 2 cases as follows:

- i. Uncertainty associated with rainfall distribution (σ_r)

$$\sigma_r = (n R)/t_R \quad (122)$$

where t_R is the duration of rainy hours (h), and n is the uncertainty of rainfall distribution (%).

- ii. Uncertainty associated with water holding capacity in addition to uncertainty associated with rainfall distribution (σ_s)

The uncertainty associated with water holding capacity is estimated by calculating standard deviation (1σ) of the hourly effective rainfall intensity among 103 cases. Then uncertainty of water holding capacity is added to the **Eq. (122)** to demonstrate the effect of both uncertainties to the uncertainty of peak runoff height.

σ_r and σ_s are represented by $r_e'(t)$ in **Eq. (121)**.

6.4 CASE STUDY

The two proposed methods are applied to a target rainfall i.e. one rainfall event occurred in Kusaki dam catchment where the total rainfall is 266 mm and the rainfall duration is 101 hours that is named case 1 rainfall.

6.4.1 ENSEMBLE METHOD

After estimating 103 values of parameter a and b , those values are used to calculate water holding capacity distribution to estimate the effective rainfall intensity for simulating runoff. The total rainfall-total rainfall loss relationship using **Eq. (6)** among 103 cases are overplotted to demonstrate the range of uncertainty of effective rainfall intensity due to uncertainty of water holding capacity (see **Figure 6.1**).

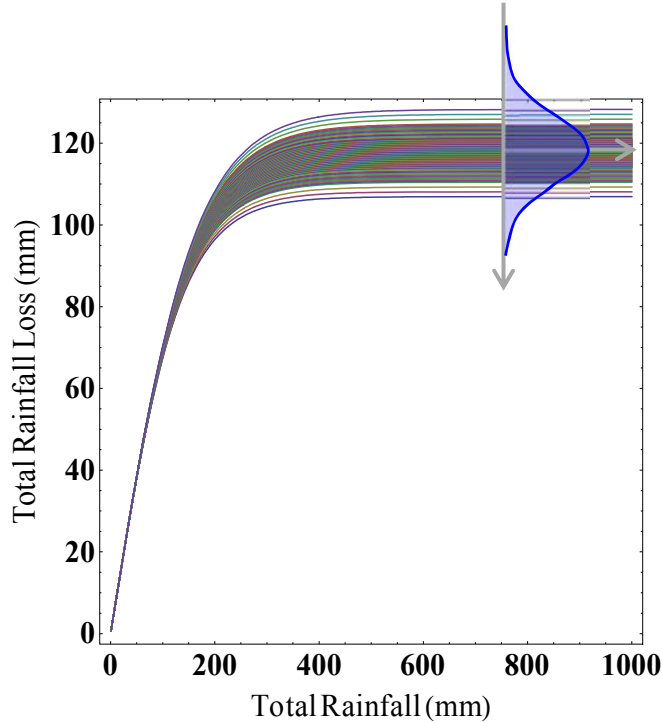


Figure 6.1. Relationship between total rainfall and total rainfall loss of 103 cases for the target rainfall.

Figure 6.1 shows that the value of total rainfall loss in the constant stage varies from 106.9 mm to 128.3 mm. This variation is mainly due to the initialization of soil moisture condition which reflects the initial water storage. If the catchment is initially wet, then more water become direct runoff when it rains. On the contrary, if the catchment is initially dry, then more water can be stored in the ground when it rains. Hence, the initial water storage affects the amount of direct runoff that causes flooding. The water holding capacity distribution is calculated by using **Eq. (9)** and the overplotting result for 103 cases is presented in **Figure 6.2**.

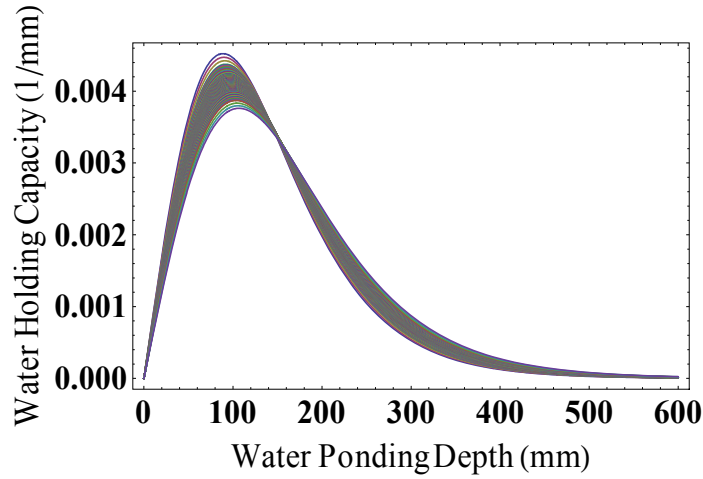


Figure 6.2. Water holding capacity distribution of 103 cases for the target rainfall.

The water holding capacity represents the capacity of a catchment to hold or to store water during a rainfall event. The area under water holding capacity distribution profile shows the outflow contribution rate. Thus, the outflow contribution rate can be obtained by integrating the water holding capacity. If the outflow contribution rate is equal to 1, it means that the initially catchment condition is saturated, hence that there is no infiltration and all rainfall becomes direct runoff. The outflow contribution rate is calculated by using **Eq. (10)** and the overplotting result for 103 cases is presented in **Figure 6.3**.

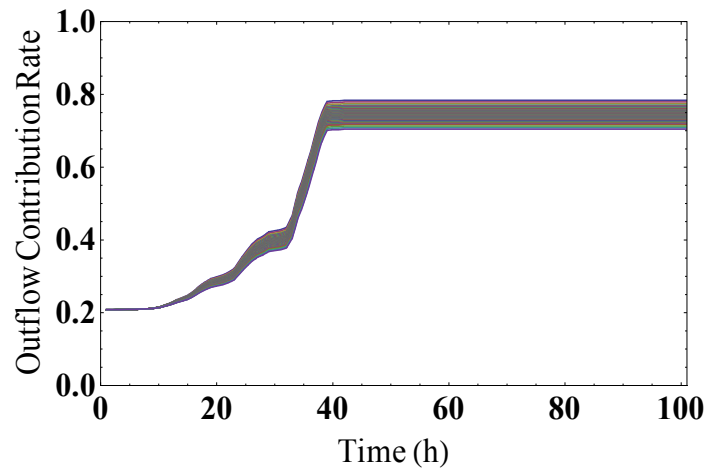


Figure 6.3. Outflow contribution rate of 103 cases for the target rainfall.

The values of outflow contribution rate among 103 cases varies from 0.70 to 0.78 (see **Figure 6.3**). After obtaining the outflow contribution rate, then the effective rainfall intensity can be calculated using **Eq. (11)**. The overplotting of effective rainfall intensity for 103 cases including its variance is presented in **Figure 6.4**.

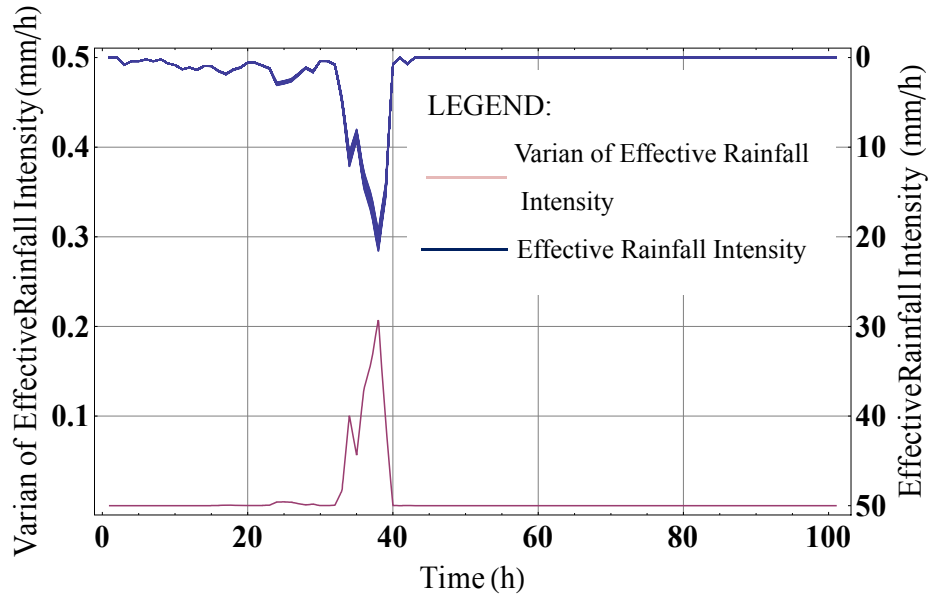


Figure 6.4. Effective rainfall intensity of 103 cases for the target rainfall.

The maximum and minimum values of effective rainfall intensity during the peak time among 103 cases are 21.7 mm/h and 19.3 mm/h, respectively. After obtaining the effective rainfall intensity, then the runoff height can be simulated by using **Eq. (31)**. The overplotting of simulated runoff height for 103 cases is presented in **Figure 6.5**.

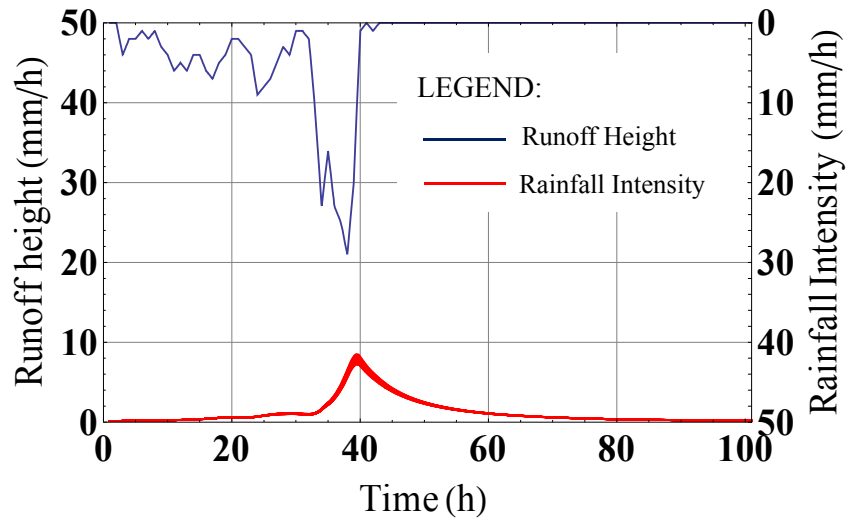


Figure 6.5. Runoff height simulation of 103 cases for the target rainfall.

The maximum and minimum values of runoff during the peak time among 103 cases are 8.5 mm/h and 7.2 mm/h, respectively.

6.4.2 STOCHASTIC DIFFERENTIAL EQUATION METHOD

In this method, two different types of uncertainties are considered to quantify the uncertainty of peak runoff height for several cases. For each case, the uncertainty associated with rainfall distribution (σ_r) is calculated for different percentage i.e. 5%, 10%, 15%, 20%, and 25%, whereas the uncertainty associated with water holding capacity (σ_s) remains the same. σ_r is calculated by using Eq. (122), whereas σ_s is obtained by calculating the standard deviation among the effective rainfall intensity of 103 cases. Then for each case of σ_r , two cases are considered i.e. by only considering uncertainty associated with rainfall distribution and by considering both uncertainties (see **Figure 6.6**).

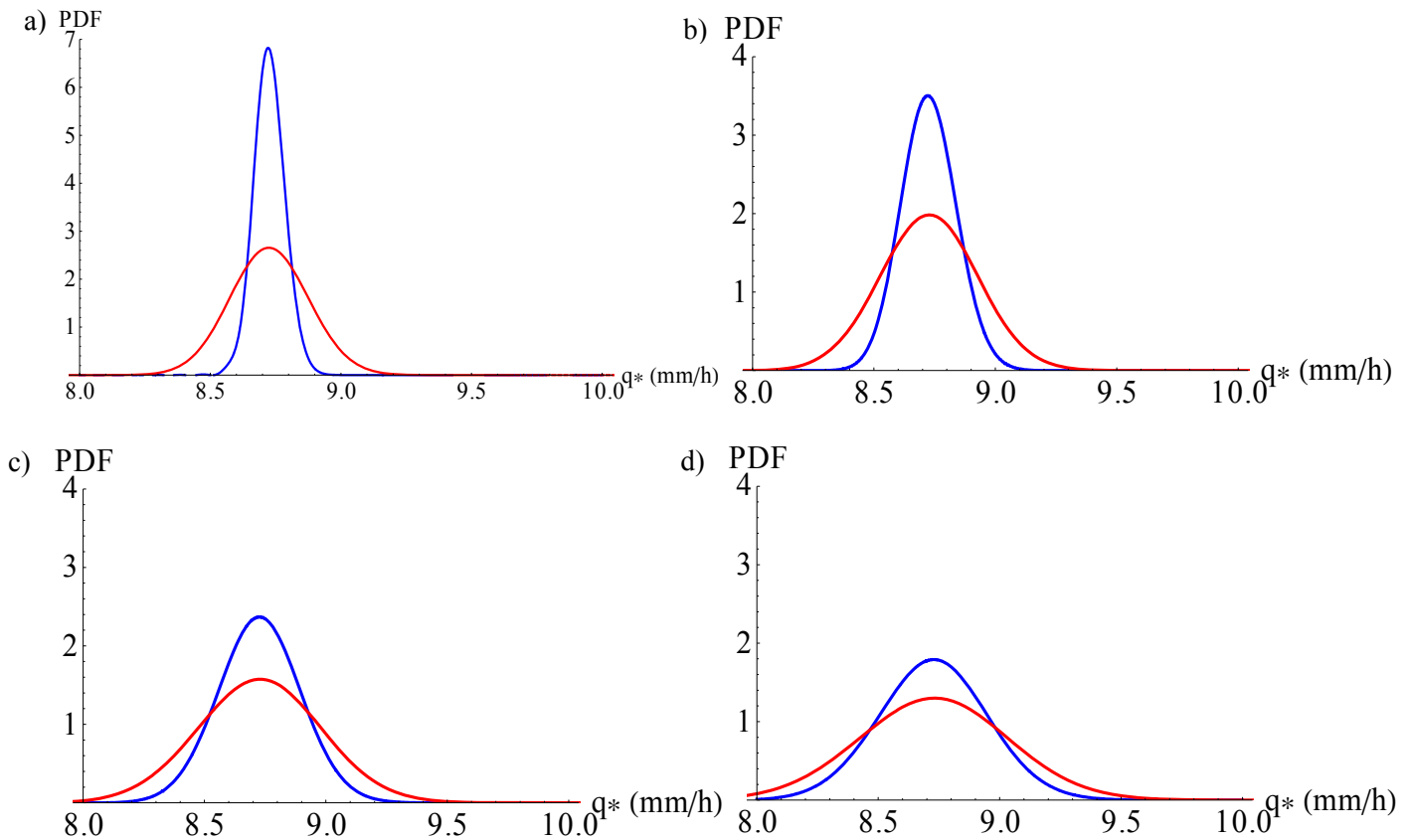


Figure 6.6. Probability density function (PDF) of peak runoff height. Blue line denotes PDF of peak runoff height based on uncertainty of rainfall distribution, and red line denotes PDF of peak runoff height based on uncertainty of rainfall distribution and uncertainty of water holding capacity. The uncertainty of rainfall distribution is considered for different cases: a) 5% b) 10% c) 15% d) 20%.

The peak runoff height uncertainty is calculated by finding the values of runoff height when the PDF value is close to zero i.e. 0.000011. The average value of peak runoff height for all of cases is 8.7 mm/h. The results are summarized in **Table 6.1**:

Table 6.1. Summary of uncertainty of peak runoff height.

No.	Uncertainty associated with rainfall pattern (%)	Uncertainty of peak runoff height associated with uncertainty due to (mm/h):		Difference between uncertainty of peak runoff height by considering both uncertainties and by considering rainfall pattern only (mm/h)
		rainfall pattern	rainfall pattern and water holding capacity	
1	5	0.6	1.5	0.9
2	10	1.2	1.9	0.7
3	15	1.6	2.4	0.8
4	20	2.1	2.9	0.8

Results show that the peak runoff height uncertainty increase with the increment of uncertainty associated with rainfall distribution, and in average, the uncertainty of water holding capacity contributes 0.8 mm/h to the uncertainty of peak runoff height. The average value of peak runoff height based on SDE method is higher than that of by the ensemble method because of the impact of uncertainty associated with rainfall distribution.

Figure 3.2 shows that majority of rainfall events occurred in the range of 0-100 mm. Thus, 10 big rainfall events having total rainfall more than 100 mm are selected for further analysis. The characteristics of each rainfall events is summarized in **Table 6.2**.

Table 6.2. Summary of selected big rainfall events occurred at Kusaki dam catchment.

No.	Rainfall event	a_0	β	m	Total Rainfall (mm)	Duration of rainfall (h)	Time of peak rainfall intensity (h)
1	9 July to 13 July 2002	0.05	0.40	0.67	363.95	97	19
2	Case 1 rainfall	0.07	0.40	0.67	266	101	38
3	8 August to 10 August 2003	0.06	0.38	0.67	240.25	50	11
4	22 August-3 September 2005	0.10	0.25	0.33	202.4	312	95
5	6-18 August 2009	0.08	0.25	0.33	187.8	312	99
6	26 July to 3 August 2011	0.10	0.39	0.64	179.7	202	122
7	19 July-2 August 2005	0.10	0.15	0.18	169.2	359	289
8	25 September - 13 October 2009	0.05	0.15	0.18	165.2	456	317
9	3 October to 8 October 2004	0.07	0.56	1.27	158.5	137	55
10	28 September - 14 October 2006	0.05	0.35	0.54	152.4	408	198

For each rainfall event, the uncertainty associated with water holding capacity (σ_s) is calculated by using the following procedure:

- Calculate the effective rainfall intensity by using 3 different values of runoff parameters of a and b i.e. a and b (average case), a and b (average + 1σ case), and a and b (average - 1σ case).
- Calculate the discrepancy between two different cases as follows:
 - σ_{s1} = Effective rainfall intensity simulated by using a and b (average case) - effective rainfall intensity simulated by using a and b (average + 1σ case).
 - σ_{s2} = Effective rainfall intensity simulated by using - effective rainfall intensity simulated by using a and b (average - 1σ case) - a and b (average case).
- Select the values of σ_{s1} and σ_{s2} at the time of peak rainfall intensity.
- Calculate the average value of σ_{s1} and σ_{s2} to obtain σ_s :

$$\sigma_s = \frac{\sigma_{s1} + \sigma_{s2}}{2} \quad (123)$$

By using σ_r for 3 different cases i.e. 10%, 15%, and 20% calculated by using **Eq. (122)**, and σ_s is calculated by using **Eq. (123)**, then the probability of uncertainty of peak runoff height for each rainfall event is simulated by using **Eq. (121)**. The sample of hydrograph is shown in **Figure 6.7**, whereas the probability of uncertainty of peak runoff height for each rainfall event is shown in **Figure 6.8**.

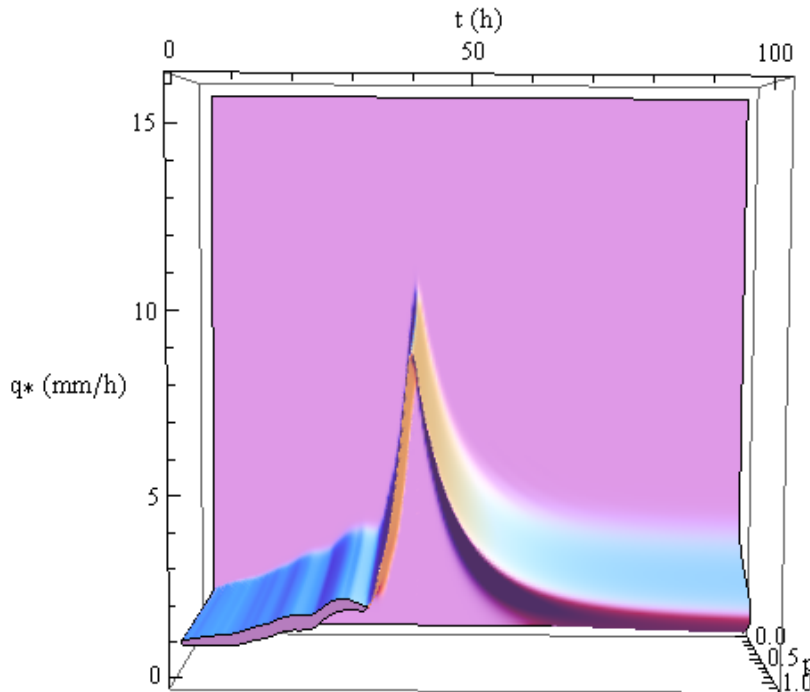
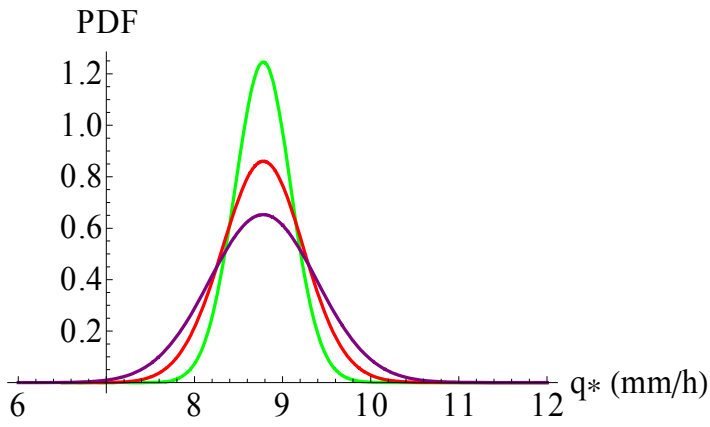
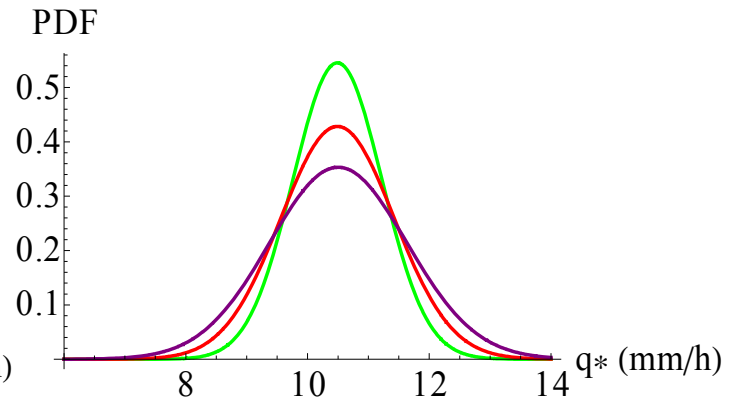


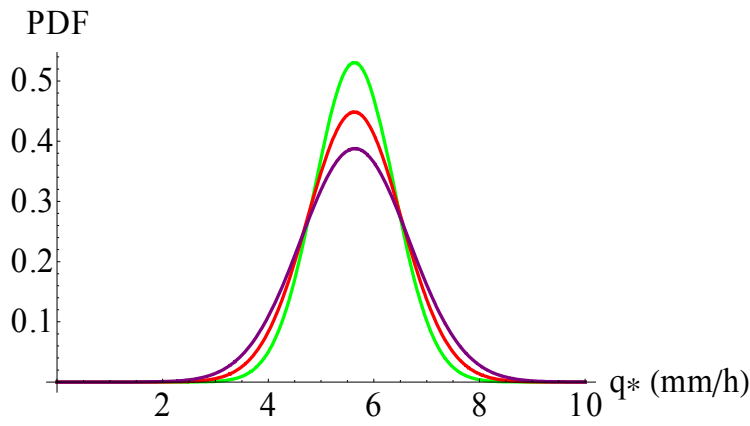
Figure 6.7. The hydrograph of case 1 rainfall.



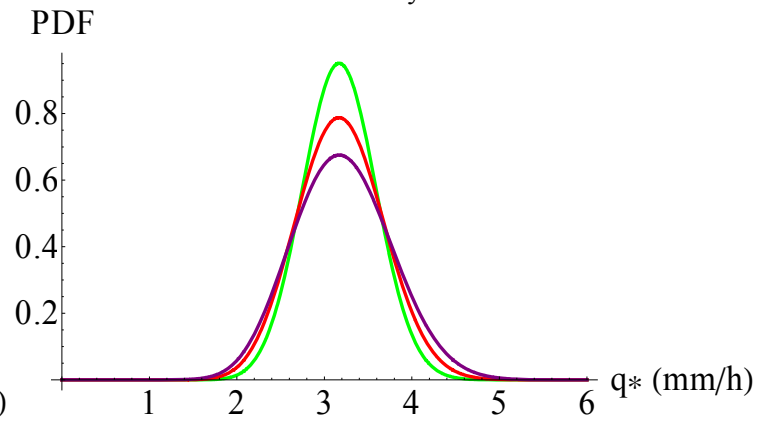
Case 1 rainfall



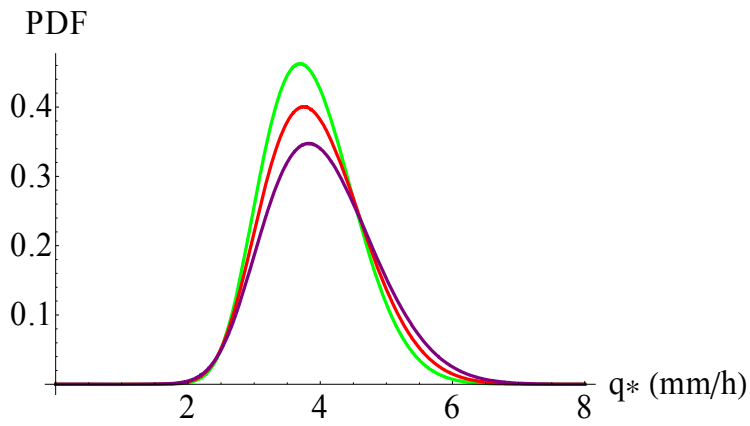
9-13 July 2002



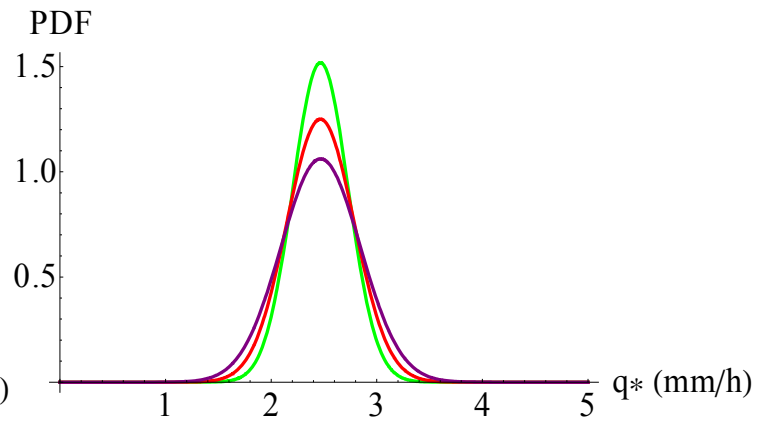
8-10 August 2003



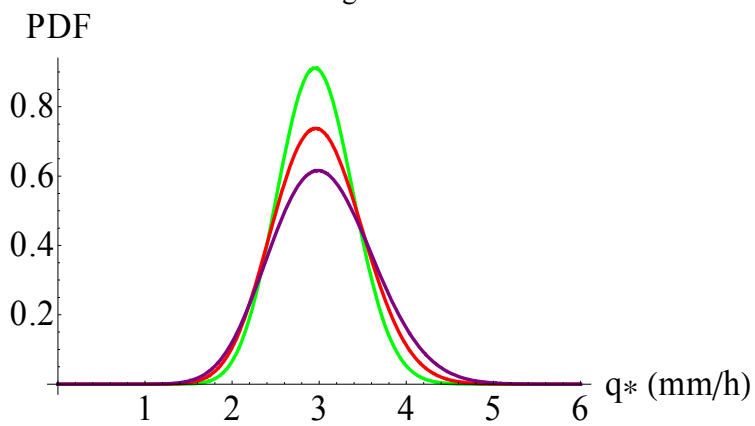
22 August – 3 September 2005



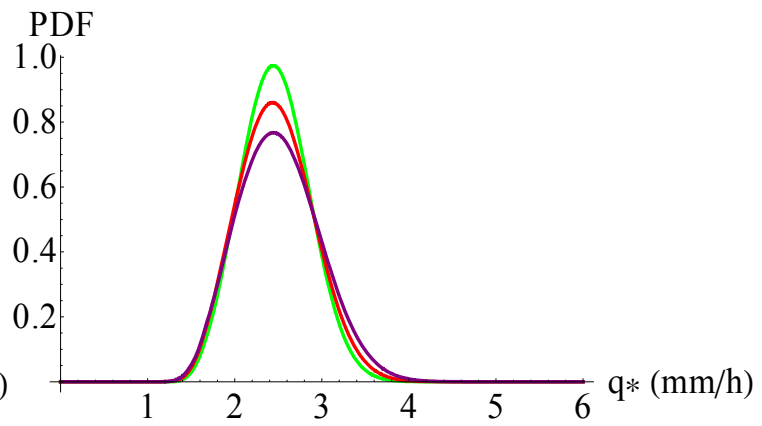
6-18 August 2009



26 July-3 August 2011



19 July- 2 August 2005



25 September – 13 October 2009

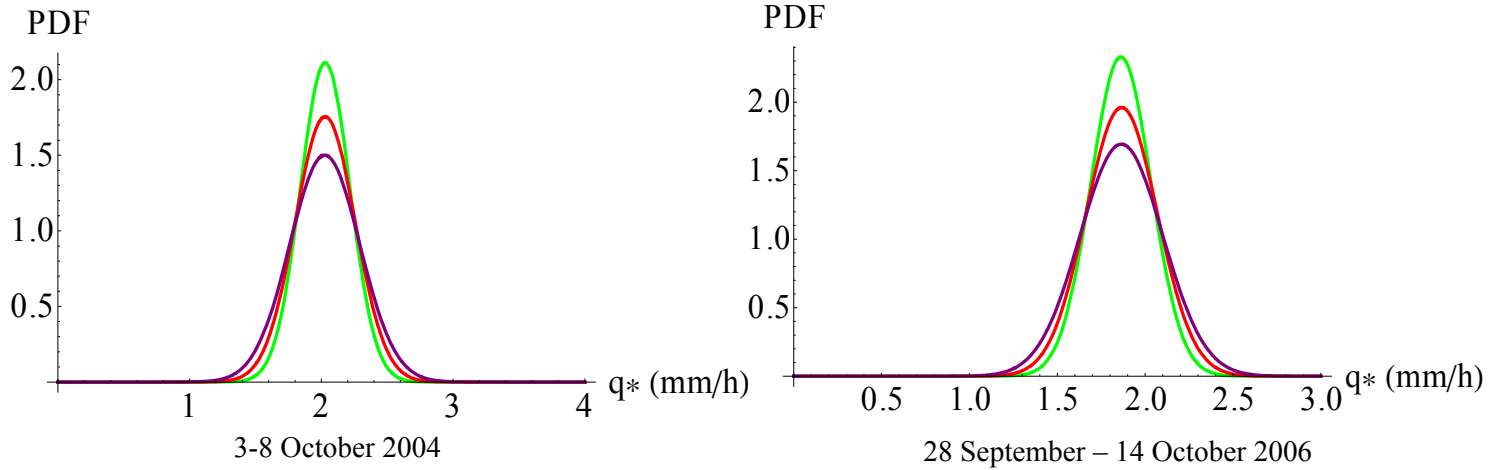


Figure 6.8. Probability density function (PDF) of peak runoff height based on uncertainties associated with uncertainty of rainfall distribution and water holding capacity. The uncertainty associated with rainfall distribution is simulated based on 3 different cases i.e. 10% (green line), 15% (red line), and 20% (purple line).

Figure 6.8 shows that the bigger uncertainty associated with rainfall distribution results in higher uncertainty of peak runoff height. The values of uncertainty of peak runoff height is summarized in **Table 6.3**.

Table 6.3. Summary of uncertainty of peak runoff height associated with uncertainties of rainfall distribution and water holding capacity.

No.	Rainfall event	a0	β	m	Total Rainfall (mm)	Duration of rainfall (h)	Time of peak rainfall intensity (h)	Uncertainty of peak runoff height associated with uncertainties of rainfall distribution and water holding capacity (mm/h)		
								$\sigma_r = 10\% + \sigma_s$	$\sigma_r = 15\% + \sigma_s$	$\sigma_r = 20\% + \sigma_s$
1	9 July to 13 July 2002	0.05	0.40	0.67	363.95	97	19	6.77	8.62	10.5
2	Case 1 rainfall	0.07	0.40	0.67	266	101	38	3	4.26	5.61
3	8 August to 10 August 2003	0.06	0.38	0.67	240.25	50	11	7.02	8.29	9.4
4	22 August-3 September 2005	0.10	0.25	0.33	202.4	312	95	3.71	4.29	5.14
5	6-18 August 2009	0.08	0.25	0.33	187.8	312	99	6.07	6.58	7.24
6	26 July to 3 August 2011	0.10	0.39	0.64	179.7	202	122	2.58	3.1	3.47
7	19 July-2 August 2005	0.10	0.15	0.18	169.2	359	289	3.87	4.51	5.13
8	25 September - 13 October 2009	0.05	0.15	0.18	165.2	456	317	3.56	3.65	3.87
9	3 October to 8 October 2004	0.07	0.56	1.27	158.5	137	55	1.88	2.25	2.6
10	28 September - 14 October 2006	0.05	0.35	0.54	152.4	408	198	1.7	2	2.3

The values of uncertainty of peak runoff height in **Table 6.3** are multiplied by catchment area to obtain the values of uncertainty of peak runoff (see **Table 6.4**).

Table 6.4. Summary of uncertainty of peak runoff associated with uncertainties of rainfall distribution and water holding capacity.

No.	Rainfall event	a0	β	m	Total Rainfall (mm)	Duration of rainfall (h)	Time of peak rainfall intensity (h)	Uncertainty of peak runoff associated with uncertainties of water holding capacity and rainfall distribution (m^3/s)		
								$\sigma_r = 10\% + \sigma_s$	$\sigma_r = 15\% + \sigma_s$	$\sigma_r = 20\% + \sigma_s$
1	9 July to 13 July 2002	0.05	0.40	0.67	363.95	97	19	496.47	632.13	770.00
2	Case 1 rainfall	0.07	0.40	0.67	266	101	38	220.73	312.40	411.40
3	8 August to 10 August 2003	0.06	0.38	0.67	240.25	50	11	514.80	607.93	689.33
4	22 August-3 September 2005	0.10	0.25	0.33	202.4	312	95	272.07	314.60	376.93
5	6-18 August 2009	0.08	0.25	0.33	187.8	312	99	445.13	482.53	530.93
6	26 July to 3 August 2011	0.10	0.39	0.64	179.7	202	122	189.20	227.33	254.47
7	19 July-2 August 2005	0.10	0.15	0.18	169.2	359	289	283.80	330.73	376.20
8	25 September - 13 October 2009	0.05	0.15	0.18	165.2	456	317	261.07	267.67	283.80
9	3 October to 8 October 2004	0.07	0.56	1.27	158.5	137	55	137.87	165.00	190.67
10	28 September - 14 October 2006	0.05	0.35	0.54	152.4	408	198	124.67	146.67	168.67

The uncertainty of peak runoff shown in **Table 6.4** is based on both uncertainties. To know which uncertainty is more dominant to the uncertainty of peak runoff, then the uncertainty of peak runoff is simulated by using each uncertainties separately. For this purpose, the number of rainfall events are increased to improve the accuracy of the result. Thus, big rainfall events occurred in 11 catchments are selected for analysis (see **Table 6.5**). The shape of hyetograph of each rainfall events is analyzed based on the occurrence of time to peak of rainfall intensity (t_p), whether it comes earlier or later than half of total duration of rainfall (t_r), or whether it comes around half of total duration of rainfall (t_r) (see **Figure 6.9**).

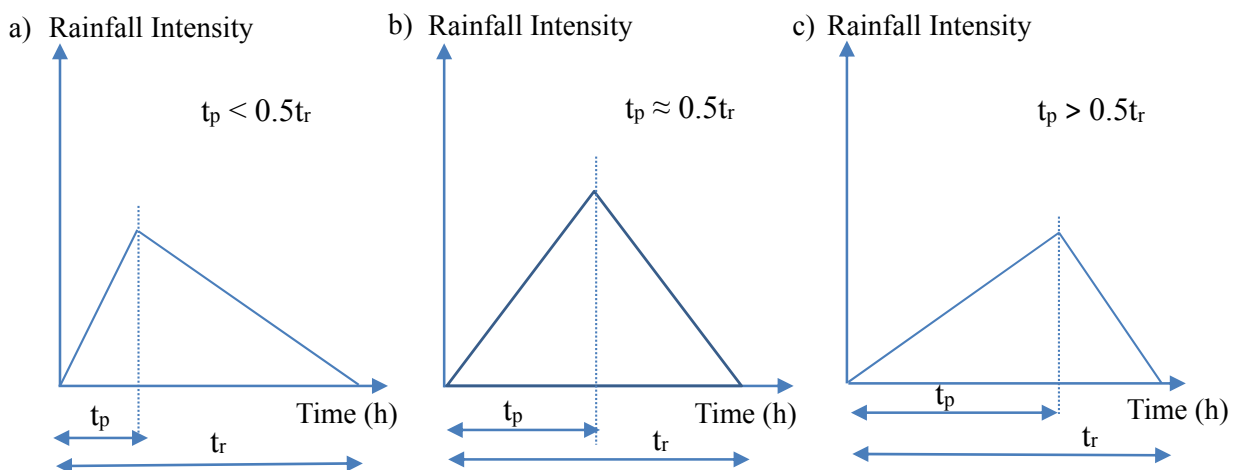


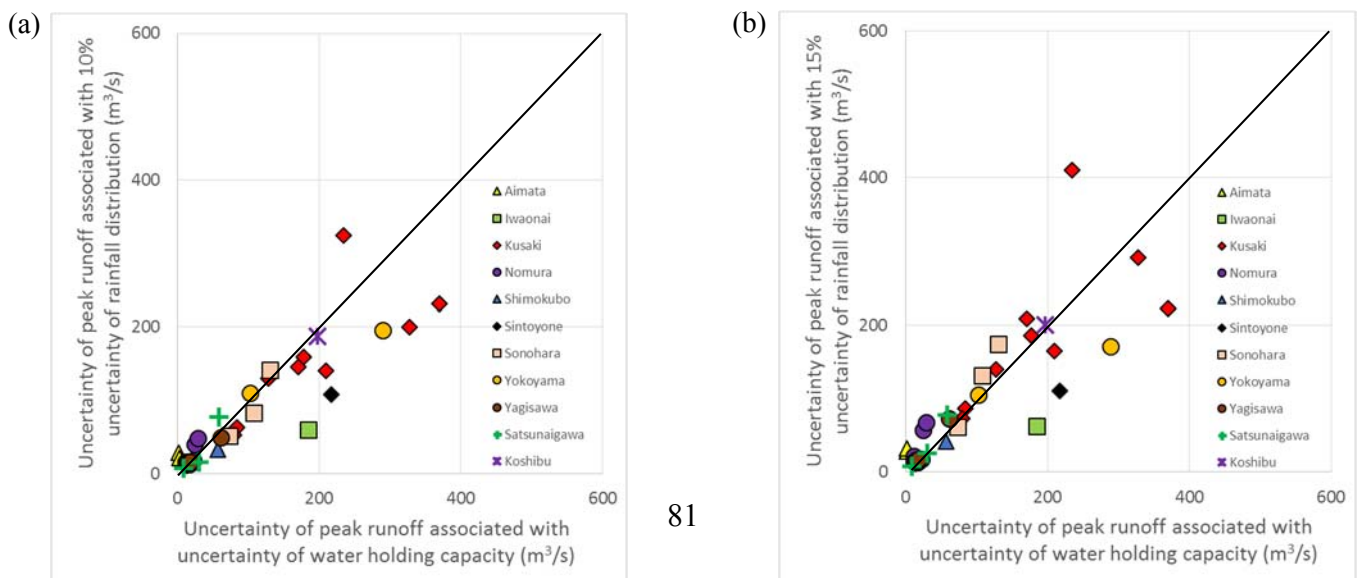
Figure 6.9. Shape of hyetograph is classified into 3 types of triangle a) triangle with peak come earlier b) isosceles triangle c) triangle with peak come later.

The summary of shape of hyetograph for each rainfall events is shown in **Table 6.5**.

Table 6.5. Summary of uncertainty of peak runoff based on each uncertainties and its shape of hyetograph

No.	Rainfall event	a0	β	m	Total Rainfall (mm)	Duration of rainfall (h)	Time of peak rainfall intensity (h)	Ratio of time to peak and duration of rainfall	Shape of hyetograph	Uncertainty of peak runoff associated with uncertainty of water holding capacity (m^3/s)	Uncertainty of peak runoff associated with uncertainty of rainfall distribution (m^3/s)			Catchment
											10%	15%	20%	
1	9-13 July 2002	0.02	0.43	0.75	257.90	98	33	0.3	Peak come earlier	28.3	21.4	31.7	39.0	Aimata
2	2-11 September 2011	0.03	0.39	0.64	225.60	225	4	0.0	Peak come earlier	184.2	60.1	62.6	67.8	Iwaonai
3	3 October to 8 October 2004	0.07	0.56	1.27	158.50	137	55	0.4	Isosceles	83.6	63.1	86.5	112.9	Kusaki
4	28 September - 14 October 2006	0.05	0.35	0.54	152.40	408	198	0.5	Isosceles	79.2	52.8	73.3	96.8	Kusaki
5	6-18 August 2009	0.08	0.25	0.33	187.80	312	99	0.3	Peak come earlier	369.6	231.7	222.2	265.5	Kusaki
6	8-10 August 2003	0.06	0.38	0.67	240.25	50	11	0.2	Peak come earlier	327.1	200.2	291.1	385.0	Kusaki
7	9-13 July 2002	0.05	0.40	0.67	363.95	97	19	0.2	Peak come earlier	234.7	324.1	409.9	544.1	Kusaki
8	22 August-3 September 2005	0.10	0.25	0.33	202.40	312	95	0.3	Peak come earlier	177.5	159.1	185.5	233.2	Kusaki
9	25 September - 13 October 2009	0.05	0.15	0.18	165.20	456	317	0.7	Peak come later	209.7	140.8	165.0	229.5	Kusaki
10	19 July-2 August 2005	0.10	0.15	0.18	169.20	359	289	0.8	Peak come later	170.9	145.2	208.3	264.0	Kusaki
11	26 July to 3 August 2011	0.10	0.39	0.64	179.70	202	122	0.6	Peak come later	127.6	129.8	140.1	166.5	Kusaki
12	4-17 September 2005	0.03	0.36	0.57	320.00	306	46	0.2	Peak come earlier	11.2	15.9	21.5	27.1	Nomura
13	30 August-3 September 2004	0.03	0.39	0.63	228.00	111	15	0.1	Peak come earlier	24.3	39.7	56.9	70.0	Nomura
14	14-17 June 2006	0.03	0.39	0.63	164.00	55	14	0.3	Peak come earlier	28.5	48.5	68.1	92.9	Nomura
15	6-10 July 2007	0.01	0.51	1.04	156.00	100	18	0.2	Peak come earlier	16.3	12.6	13.5	14.5	Nomura
16	9-15 July 2005	0.01	0.55	1.23	155.00	157	42	0.3	Peak come earlier	11.7	13.5	14.5	16.3	Nomura
17	28 September-6 October 2004	0.01	0.48	0.92	137.00	194	22	0.1	Peak come earlier	22.4	18.7	17.3	18.2	Nomura
18	11-21 July 2003	0.00	0.65	1.85	124.00	224	22	0.1	Peak come earlier	16.3	14.5	14.9	15.9	Nomura
19	13-22 July 2007	0.03	0.30	0.43	149.30	201	151	0.8	Peak come later	56.5	32.9	41.0	51.7	Shimokubo
20	19-27 September 2011	0.00	0.64	1.82	317.90	174	42	0.2	Peak come earlier	13.0	11.1	10.8	11.1	Sintoyone
21	20-24 June 2004	0.01	0.54	1.17	218.00	118	39	0.3	Peak come earlier	12.4	12.1	12.4	12.7	Sintoyone
22	8-9 September 2010	0.10	0.31	0.44	119.70	31	9	0.3	Peak come earlier	216.7	108.0	110.8	153.8	Sintoyone
23	19-22 October 2004	0.01	0.53	1.11	208.00	88	38	0.4	Isosceles	13.3	12.1	11.1	14.2	Sintoyone
24	10-13 July 2002	0.04	0.57	1.34	182.30	85	17	0.2	Peak come earlier	130.3	141.0	174.2	228.8	Sonohara
25	19-28 October 2004	0.03	0.41	0.68	136.00	211	99	0.5	Isosceles	72.9	51.8	61.5	73.5	Sonohara
26	8-11 August 2003	0.04	0.52	1.10	133.40	72	12	0.2	Peak come earlier	107.2	82.5	131.0	148.4	Sonohara
27	8-13 August 2003	0.04	0.59	1.44	275.80	137	18	0.1	Peak come earlier	289.1	196.2	170.8	236.2	Yokoyama
28	4-16 September 2005	0.04	0.60	1.52	241.50	291	62	0.2	Peak come earlier	101.8	110.5	105.5	119.2	Yokoyama
29	9-13 July 2002	0.01	0.59	1.41	178.70	93	26	0.3	Peak come earlier	14.4	16.3	16.4	15.9	Yagisawa
30	17-19 July 2004	0.04	0.34	0.52	115.00	33	8	0.2	Peak come earlier	61.3	49.7	72.4	95.7	Yagisawa
31	25-29 July 2005	0.01	0.49	0.96	104.80	99	92	0.9	Peak come later	17.3	16.2	17.2	16.9	Yagisawa
32	7-13 September 2005	0.03	0.37	0.58	311.00	157	19	0.1	Peak come earlier	58.3	77.4	77.8	78.3	Satsunaigawa
33	1-5 October 2002	0.02	0.44	0.79	191.00	88	17	0.2	Peak come earlier	30.0	15.7	25.1	29.9	Satsunaigawa
34	18-21 July 2009	0.01	0.57	1.35	132.40	81	24	0.3	Peak come earlier	8.6	7.6	7.8	8.0	Satsunaigawa
35	28 September-4 October 2002	0.10	0.29	0.41	113.00	153	88	0.6	Peak come later	196.9	187.4	200.5	238.3	Koshibu

Based on the results shown in **Table 6.5**, the uncertainty of peak runoff associated with uncertainty of water holding capacity is plotted against uncertainty of peak runoff associated with uncertainty of rainfall distribution (see **Figure 6.10**).



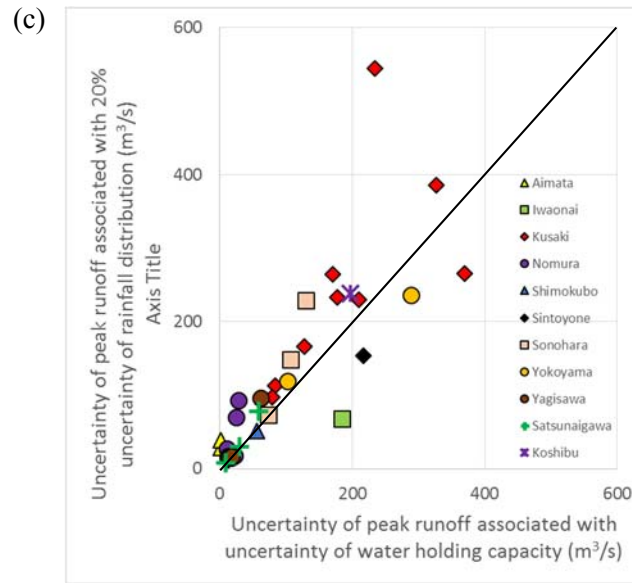


Figure 6.10. 1 to 1 plot between uncertainty of peak runoff associated with uncertainty of water holding capacity and uncertainty of peak runoff associated with uncertainty of rainfall distribution for different uncertainty (a) 10%, (b) 15%, and (c) 20%.

From **Figure 6.10**, it can be seen that the uncertainty of peak runoff associated with water holding capacity tends to be more dominant than uncertainty of peak runoff associated with 10% uncertainty of rainfall distribution (see **Figure 6.10 (a)**). **Figure 6.10 (c)** showed that quantification of uncertainty of peak runoff associated with 20% uncertainty of rainfall distribution is more dominant than that of associated with uncertainty of water holding capacity.

6.5 SUMMARY

The new findings found on this chapter can be summarized as follows:

- Two different methods namely ensemble method and stochastic differential equation are proposed to quantify the uncertainty of peak runoff. Results showed that the uncertainty of peak runoff increased with the increment of uncertainty associated with rainfall distribution.
- Uncertainty of water holding capacity needs to be included in the quantification of peak runoff uncertainty.
- Among 35 big rainfall events occurred in 11 catchments, majority of rainfall events having hyetograph with peak come earlier.
- Uncertainty associated with water holding capacity contributes more to the quantification of uncertainty of peak runoff when the uncertainty of rainfall distribution is 10%. And it contributes less to the quantification of uncertainty of peak runoff when the uncertainty of rainfall distribution is 20%.

Chapter 7. CONCLUSIONS

7.1. CONCLUSIONS

Simulating runoff with higher accuracy is one of the main research target in the hydrology. This study is using total rainfall and total rainfall loss relationship to estimate effective rainfall intensity to simulate runoff. Runoff parameters namely parameter a and parameter b describing relationship between the total rainfall and total rainfall loss can be used to estimate effective rainfall intensity for simulating runoff, if those runoff parameters are obtained when the target catchment has already experienced saturation conditions. Such saturation conditions are characterized by the constant stage after the linear stage in the empirical relationship, which can be described by a two-parameter tanh-type curve.

This study found that one of the two parameters describing that relationship, the parameter a , is the maximum height of the tanh-curve, and can be interpreted as the actual potential catchment storage. Therefore, parameter a is named total rainfall loss under saturation condition because it can be obtained when the catchment has ever experienced saturation condition. From the 36 studied catchments, 23 catchments experienced such saturation conditions. Results from these 23 catchments show that the average value of total rainfall that causes saturation conditions with a constant-stage tanh-type curve is 283 mm. A catchment has experienced saturation conditions if heavy rainfall events with small total rainfall loss have already occurred in that catchment.

As many catchments have not yet experienced such conditions, two methods namely localized gradient method and inverse method are proposed to investigate the possibility of using a limited data set (a data set of rainfall amounts that do not cause the saturation conditions) for estimating runoff parameters at the constant stage (saturated conditions). For the first method, the localized gradient method, information from catchments that experienced already saturation conditions is required to estimate runoff parameters at the constant stage. Moreover, a full available data set for the selected total rainfall is used to obtain the new values of a and b . For the second method, the inverse method, only observations from the target catchment are used for estimating the runoff parameters at the constant stage. Furthermore, only a few rainfall events, randomly chosen for each range of total rainfall, are considered. So, for this second method, it is not necessary to use the full available data set.

Each of the two proposed methods has advantages and disadvantages. The inverse method does not require information from so many catchments, uses fewer rainfall events, and does not require to calibrate a relationship between the a and b parameters. Nevertheless, to define which range of rainfall event is acceptable for use, trial and error is required. For the localized gradient method, the range of rainfall events to be used is fixed at 0–200 mm, but it requires information from many catchments with a constant-stage tanh-type curve to produce empirical equations for estimating the runoff parameters at the constant stage.

Importantly, results show that the similar values of total rainfall have different values of total rainfall loss. It is interpreted that there is uncertainty due to initialization of soil moisture condition, whether initially catchment is wet or dry before the target rainfall event occurred. Rainfall as input

data also contains uncertainty. The rainfall data used in this study is from the rain gauges measurement, where rain gauges based rainfall intensity measurements can be biased by factors like wind and evaporation in the range of 10-20%. Those uncertainties lead to the uncertainty of runoff. Thus, the main objective of this study is to quantify the uncertainty of runoff associated with uncertainties of rainfall distribution and water holding capacity by proposing two methods namely ensemble method and stochastic differential equation method. Results show that peak runoff height uncertainty increase with the increment of uncertainty associated with rainfall pattern, and uncertainty of water holding capacity needs to be included in the quantification of the uncertainty of peak runoff height.

By analyzing the shape of hyetograph for 35 big rainfall events occurred at 11 catchments, it can be concluded that majority of rainfall events occurred with peak rainfall come earlier than half of the total duration of rainfall. The uncertainty associated with water holding capacity tends to be more dominant than uncertainty associated with rainfall distribution having 10% uncertainty in quantification of uncertainty of peak runoff, and it became less dominant when the uncertainty associated with rainfall distribution having 20% uncertainty.

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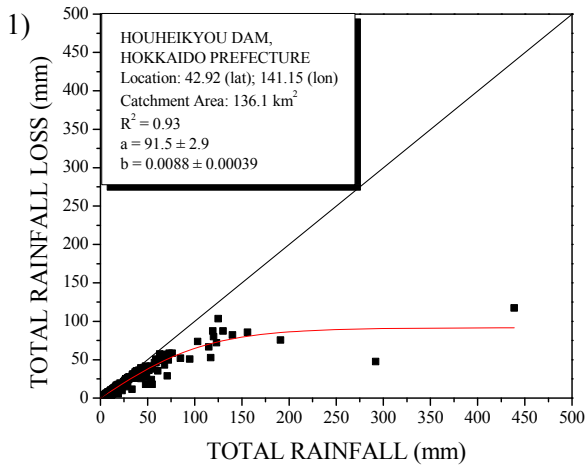
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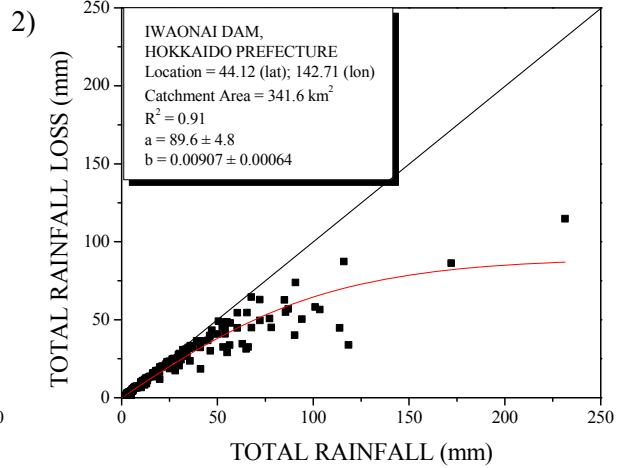
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APPENDICES

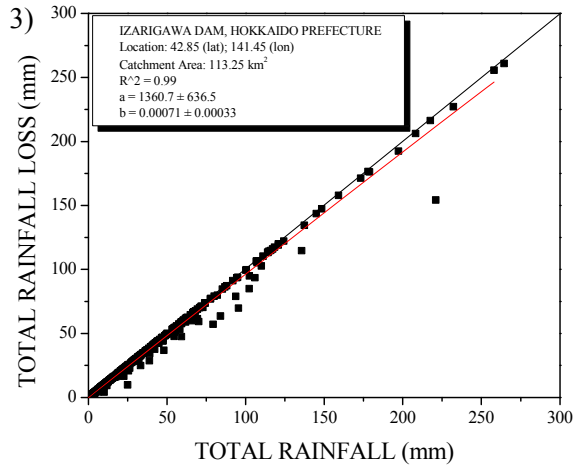
Appendix 1. Total rainfall and total rainfall loss relationship of 36 catchments



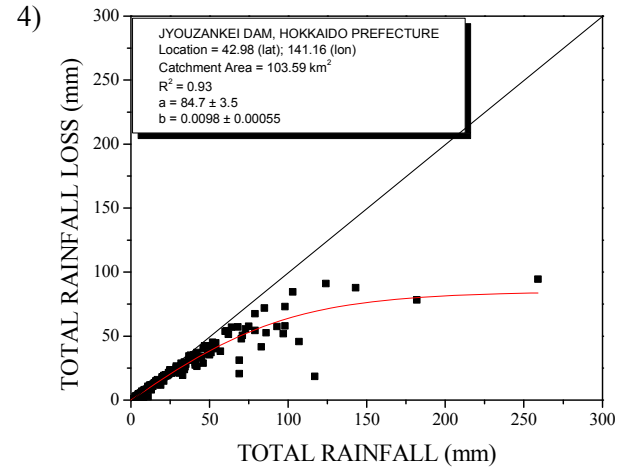
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Houheikyou dam catchment in Hokkaido Prefecture



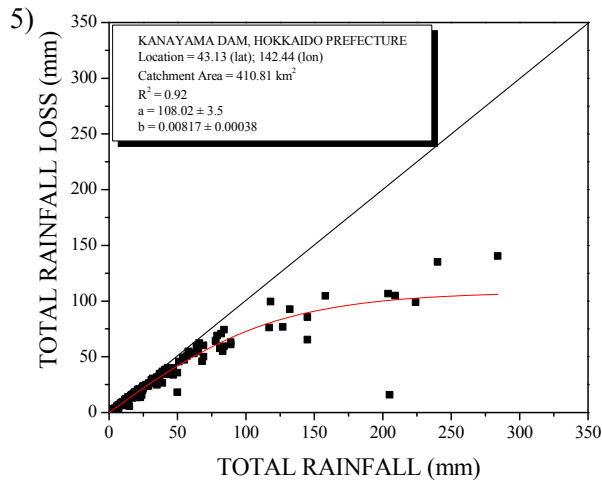
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Iwaonai dam catchment in Hokkaido Prefecture



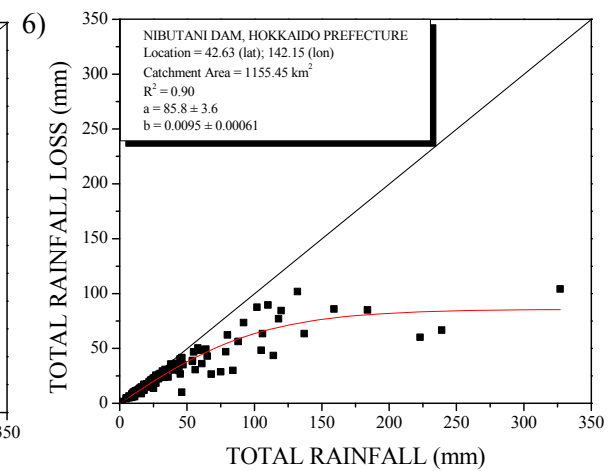
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Izarigawa dam catchment in Hokkaido Prefecture



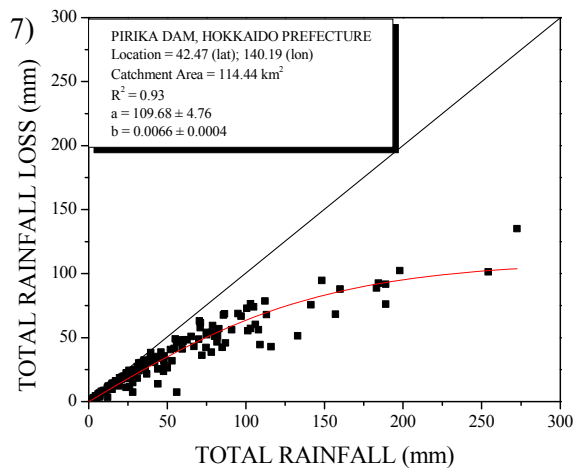
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Jyozankei dam catchment in Hokkaido Prefecture



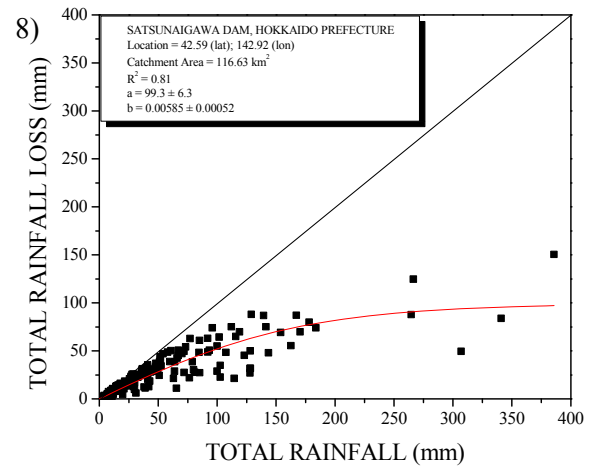
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Kanayama dam catchment in Hokkaido Prefecture



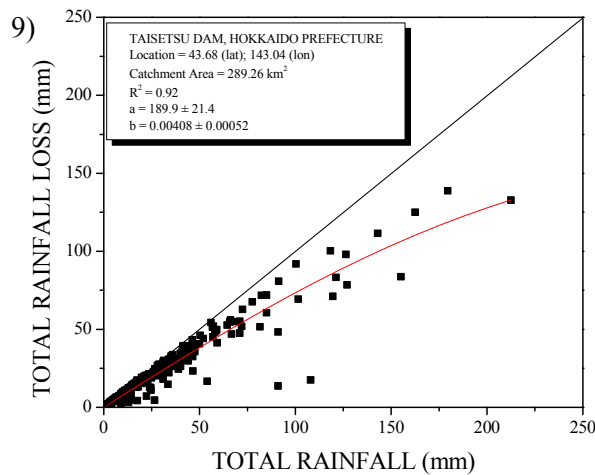
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Nibutani dam catchment in Hokkaido Prefecture



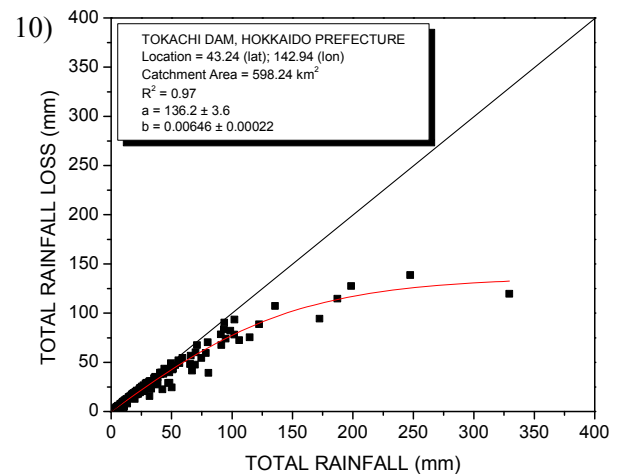
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Pirika dam catchment in Hokkaido Prefecture



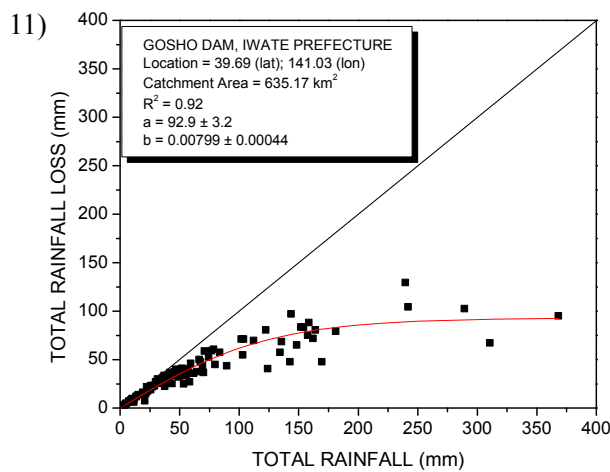
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Satsunaigawa dam catchment in Hokkaido Prefecture



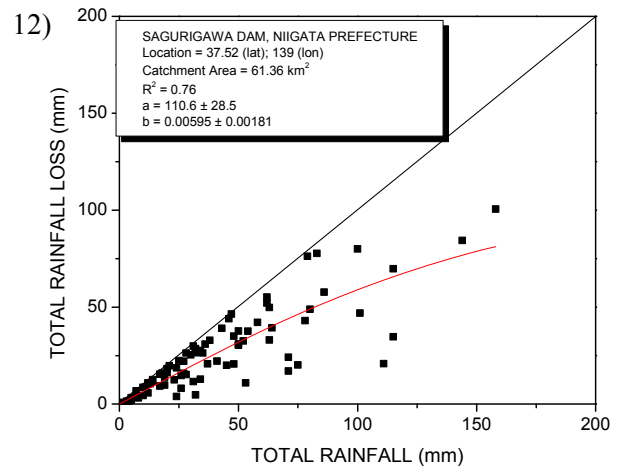
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Taisetsu dam catchment in Hokkaido Prefecture



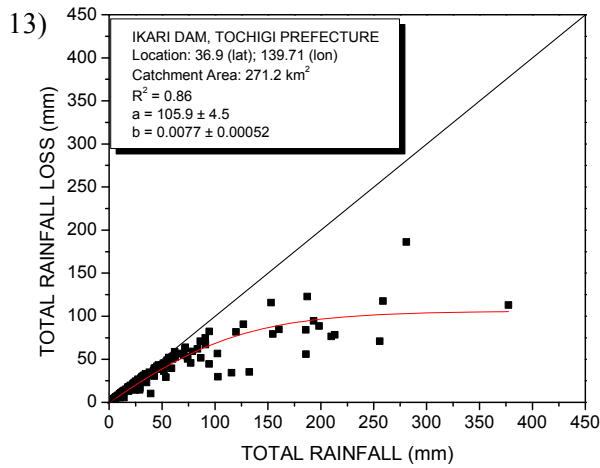
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Tokachi dam catchment in Hokkaido Prefecture



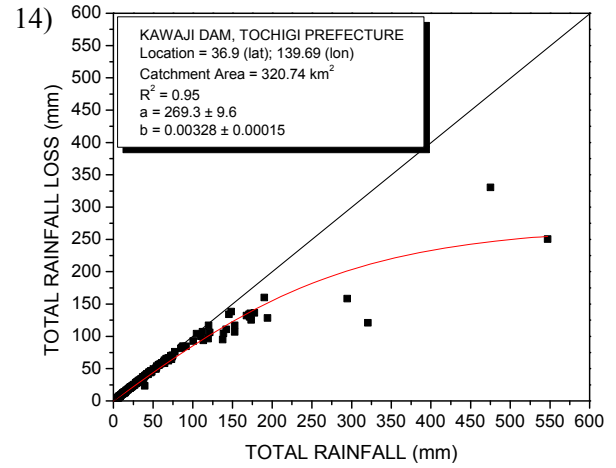
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Gosho dam catchment in Iwate Prefecture



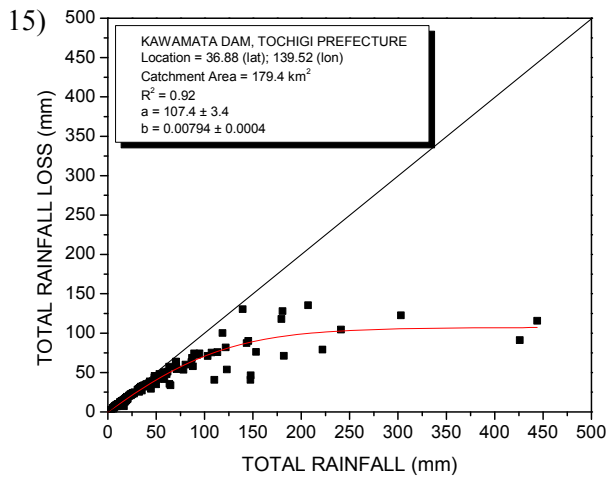
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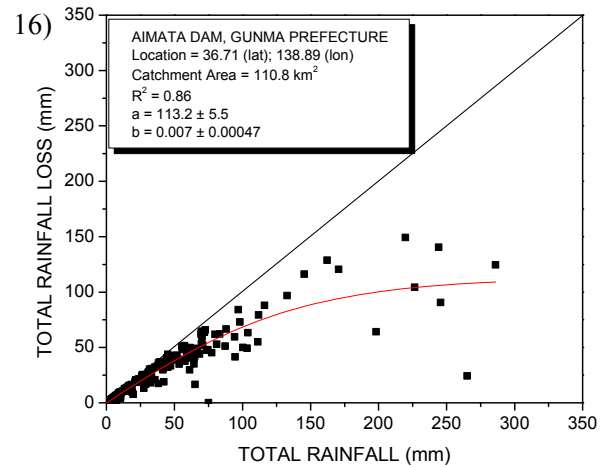
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Ikari dam catchment in Tochigi Prefecture



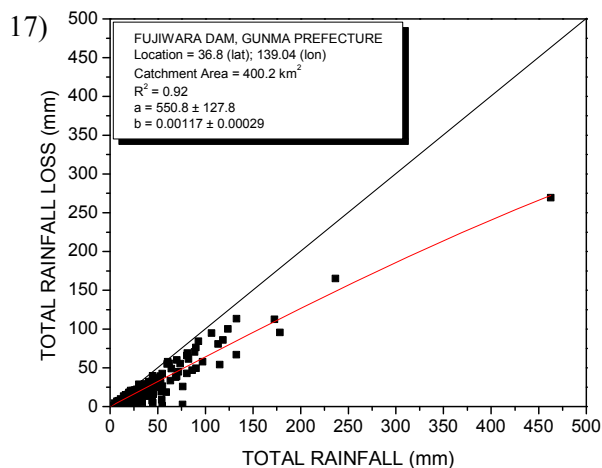
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Kawaji dam catchment in Tochigi Prefecture



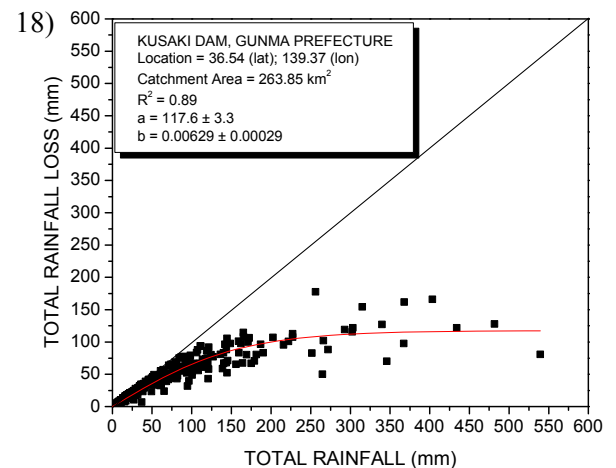
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Kawamata dam catchment in Tochigi Prefecture



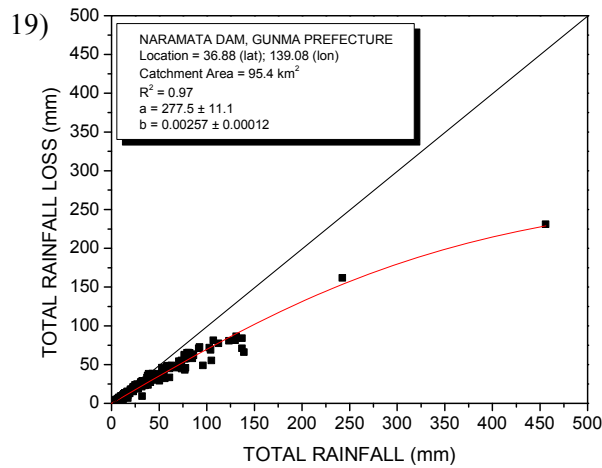
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Aimata dam catchment in Gunma Prefecture



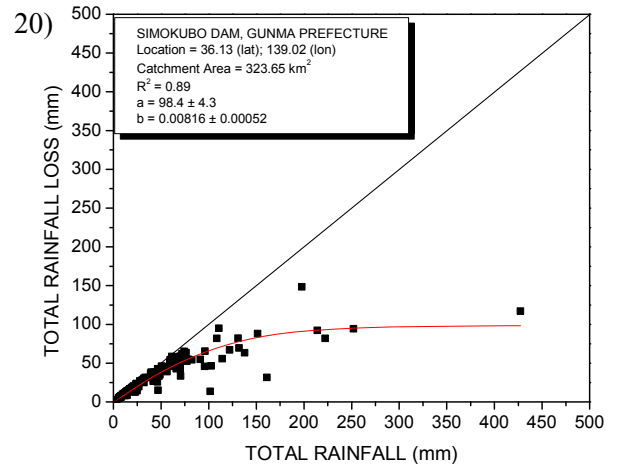
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Fujiwara dam catchment in Gunma Prefecture



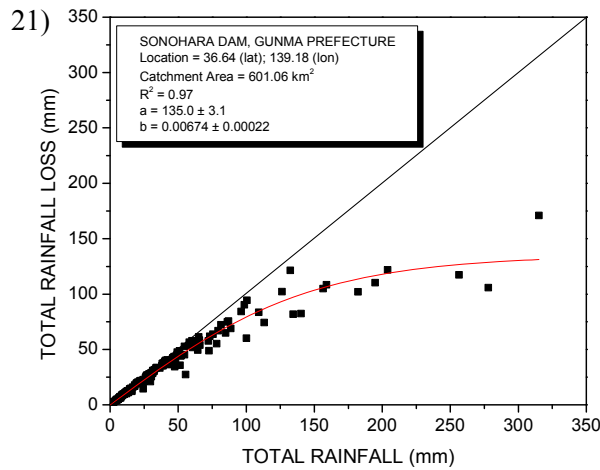
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Kusaki dam catchment in Gunma Prefecture



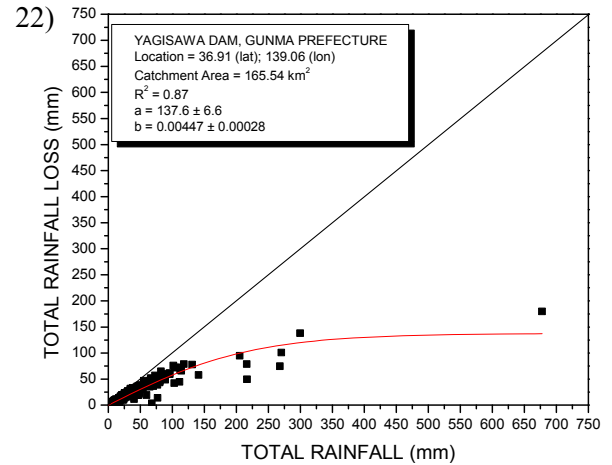
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Naramata dam catchment in Gunma Prefecture



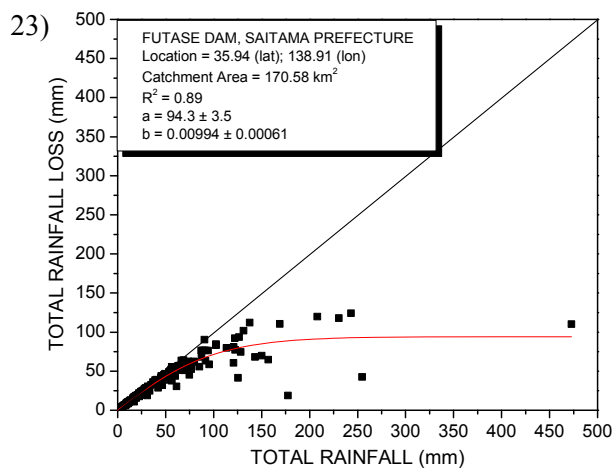
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Simokubo dam catchment in Gunma Prefecture



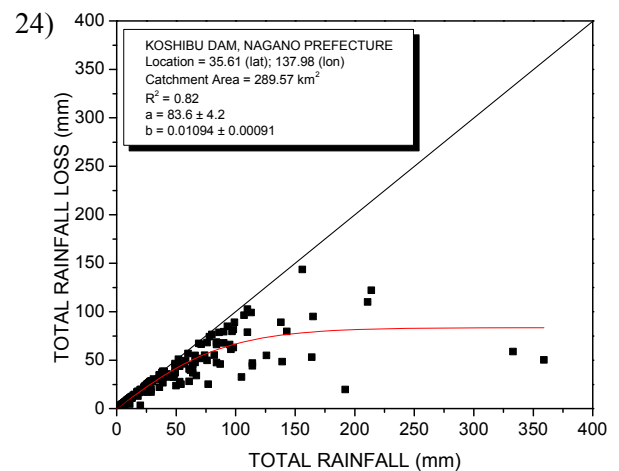
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Sonohara dam catchment in Gunma Prefecture



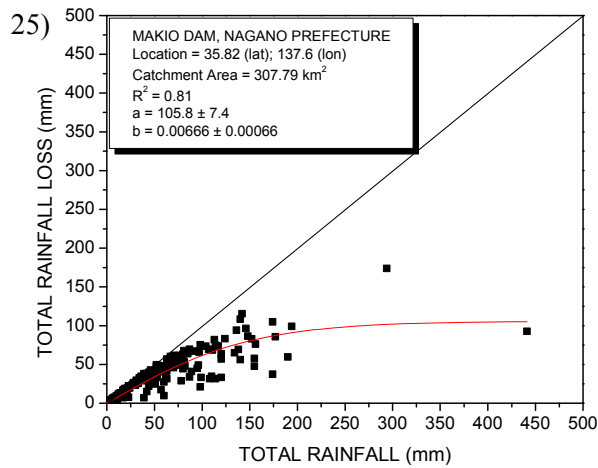
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Yagisawa dam catchment in Gunma Prefecture



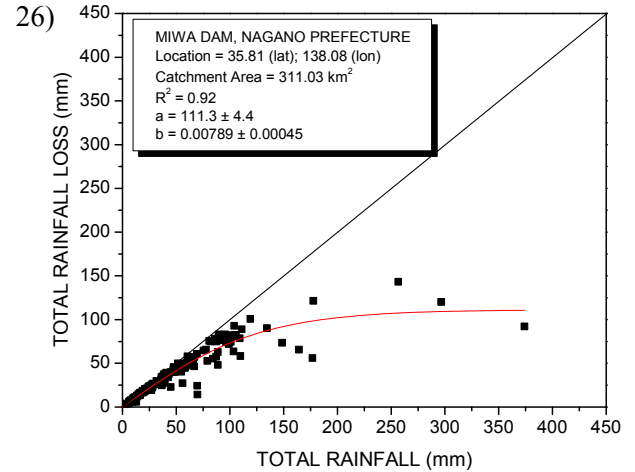
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Futase dam catchment in Saitama Prefecture



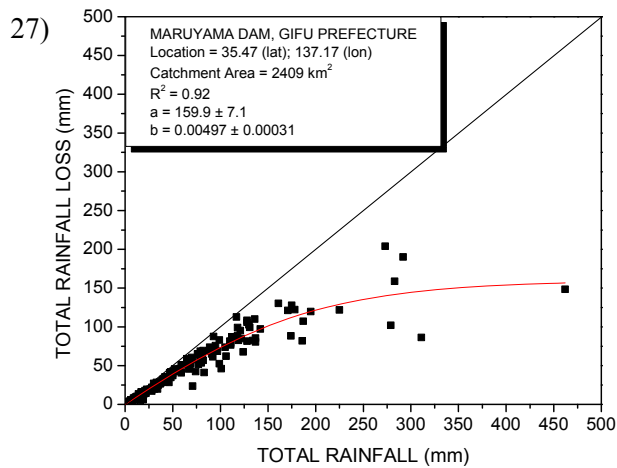
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Koshibu dam catchment in Nagano Prefecture



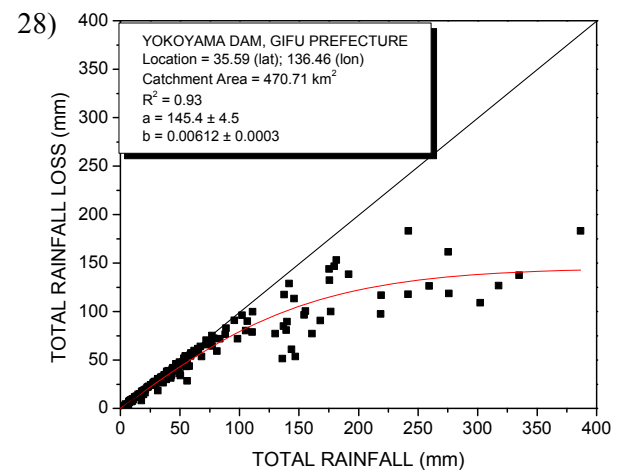
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Makio dam catchment in Nagano Prefecture



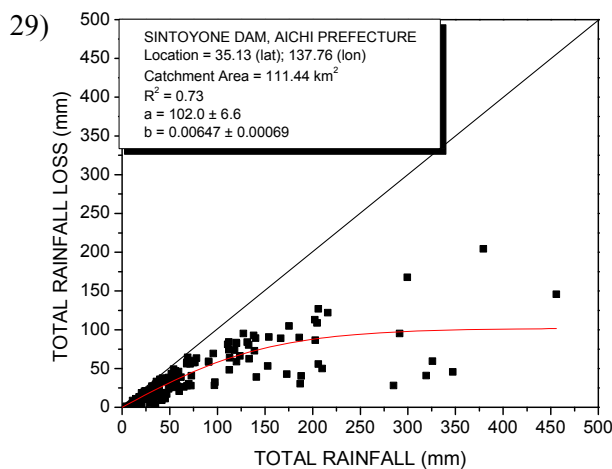
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Miwa dam catchment in Nagano Prefecture



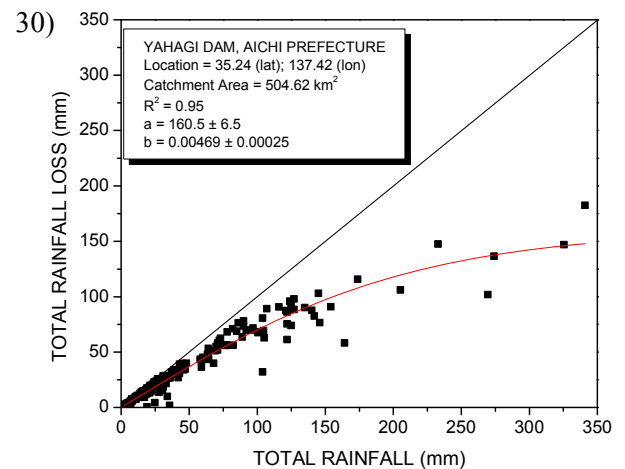
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Maruyama dam catchment in Gifu Prefecture



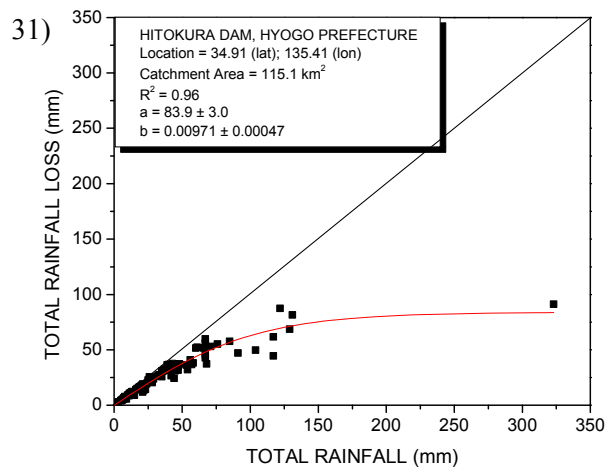
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Yokoyama dam catchment in Gifu Prefecture



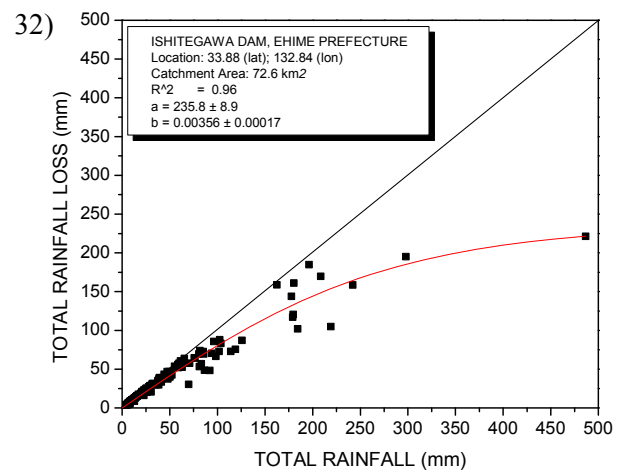
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Sintoyone dam catchment in Aichi Prefecture



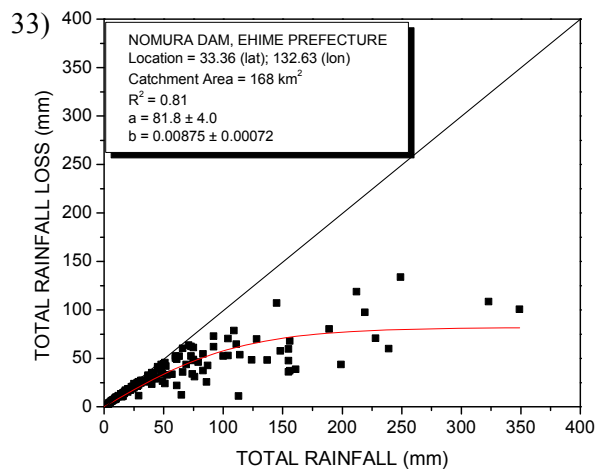
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Yahagi dam catchment in Aichi Prefecture



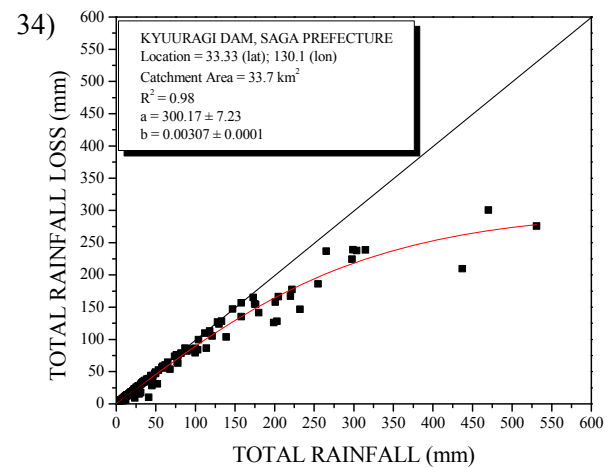
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Hitokura dam catchment in Hyogo Prefecture



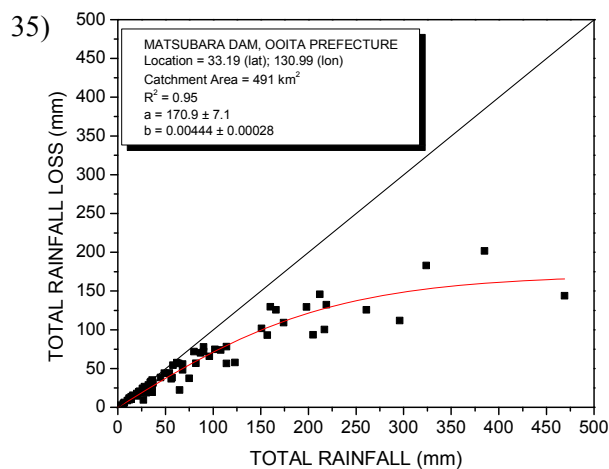
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Ishitegawa dam catchment in Ehime Prefecture



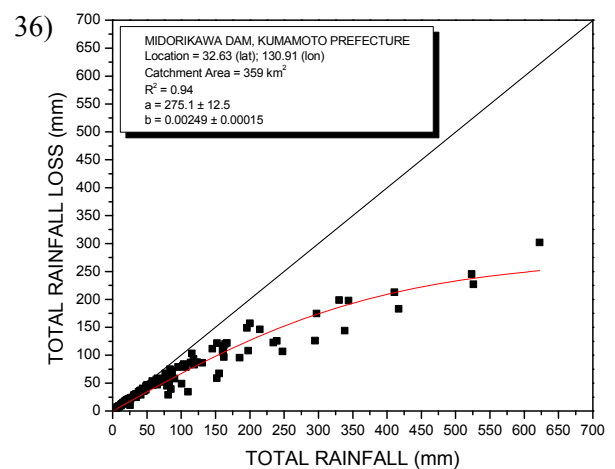
Relationship between total rainfall (mm) and total rainfall loss (mm) for the Hitokura dam catchment in Hyogo Prefecture



Relationship between total rainfall (mm) and total rainfall loss (mm) for the Kyuuragi dam catchment in Saga Prefecture



Relationship between total rainfall (mm) and total rainfall loss (mm) for the Matsubara dam catchment in Ooita Prefecture



Relationship between total rainfall (mm) and total rainfall loss (mm) for the Midorikawa dam catchment in Kumamoto Prefecture

Appendix 2. Summary of Soil Types

No.	Dam Catchment	Catchment Area (km ²)	Soil Types	Area (%)	Area (km ²)
1	Houheikyou	136.10	Dark Colored Brown Forest Soil_Soil Podsolization	67.4	91.8
			Brown Forest Soil (II)	27.3	37.2
			Podsolization Soil-Alpine Debris Soil	5.3	7.2
2	Iwaonai	341.58	Dark Colored Brown Forest Soil_Soil Podsolization	50.4	172.3
			Brown Forest Soil (II)	40.6	138.6
			Brown Forest Soil-Soil Dry Podsolization	3.0	10.1
			Podsolization Soil-Alpine Debris Soil	2.9	9.9
			Brown Forest Soil (IV)	1.6	5.6
			Coarse Brown Lowland Soil	1.1	3.9
			Brown Forest Soil (I)	0.3	1.2
3	Izarigawa	113.25	Brown Forest Soil (II)	43.5	49.3
			Black Soil (a)	39.4	44.6
			Dark Colored Brown Forest Soil_Soil Podsolization	9.3	10.5
			Brown Forest Soil_Andosol	5.2	5.9
			Coarse Brown Lowland Soil	2.5	2.8
			Extract Immature Soil Course Volcano-Brown Forest Soil	0.1	0.2
			Podsolization Soil-Alpine Debris Soil	0.1	0.1
4	Jyouzankei	103.59	Brown Forest Soil (II)	69.8	72.4
			Dark Colored Brown Forest Soil_Soil Podsolization	29.2	30.3
			Podsolization Soil-Alpine Debris Soil	0.9	1.0

No.	Dam Catchment	Catchment Area (km ²)	Soil Types	Area (%)	Area (km ²)
5	Kanayama	410.81	Brown Forest Soil (II)	35.5	145.9
			Black Soil (a)	26.4	108.5
			Brown Forest Soil_Andosol	13.0	53.3
			Brown Forest Soil (IV)	12.4	50.9
			Dark Colored Brown Forest Soil_Soil Podsolization	11.0	45.0
			Coarse Brown Lowland Soil	0.7	2.8
			Podsolization Soil-Alpine Debris Soil	0.7	2.8
			Brown Forest Soil (I)	0.3	1.4
			Light Colored Black Soil (a)	0.1	0.2
6	Nibutani	1155.45	Dark Colored Brown Forest Soil_Soil Podsolization	32.4	374.1
			Brown Forest Soil (II)	30.5	352.4
			Extract Immature Soil Course Volcano-Brown Forest Soil	14.9	171.9
			Extract Immature Coarse Grained Volcanic Soil	6.5	75.6
			Brown Forest Soil (III)	5.9	68.4
			Immature Black Soil	4.1	47.1
			Coarse Brown Lowland Soil	2.9	33.7
			Podsolization Soil-Alpine Debris Soil	2.4	28.0
			Fine Grain Gray Lowland Soil	0.1	1.4
			Extract Immature Wet Coarse Volcanic Soil	0.1	1.4
			Alpine Soil Debris-Rock Land	0.1	1.3
			Black Soil (a)	0.0	0.1
7	Pirika	114.44	Brown Forest Soil (II)	54.4	62.3
			Dark Colored Brown Forest Soil_Soil Podsolization	22.4	25.6
			Brown Forest Soil-Soil Dry Podsolization	12.5	14.3
			Brown Forest Soil (III)	8.9	10.2
			Brown Forest Soil (IV)	1.6	1.8
			Coarse Brown Lowland Soil	0.2	0.3
			Brown Forest Soil_Andosol	0.0	0.0

No.	Dam Catchment	Catchment Area (km ²)	Soil Types	Area (%)	Area (km ²)
8	Satsunaigawa	116.63	Soil Debris-Rock Land	87.2	101.6
			Alpine Soil Debris-Rock Land	7.8	9.1
			Podsolization Soil-Alpine Debris Soil	3.2	3.8
			Dark Colored Brown Forest Soil_Soil Podsolization	1.2	1.4
			Brown Forest Soil (II)	0.4	0.4
			Residual Product of Immature Soil	0.2	0.2
9	Taisetsu	289.26	Brown Forest Soil-Soil Dry Podsolization	35.9	103.8
			Dark Colored Brown Forest Soil_Soil Podsolization	32.9	95.1
			Alpine Soil Debris-Rock Land	16.9	49.0
			Podsolization Soil (I)	9.6	27.8
			Podsolization Soil-Alpine Debris Soil	3.8	11.0
			Soil Debris-Rock Land	0.5	1.4
			Podsolization Soil_II	0.2	0.6
			High Peat Soil	0.1	0.4
10	Tokachi	598.24	Brown Forest Soil (II)	61.9	370.1
			Dark Colored Brown Forest Soil_Soil Podsolization	20.1	120.2
			Podsolization Soil_II	10.9	65.3
			Podsolization Soil-Alpine Debris Soil	1.5	8.8
			Soil Debris-Rock Land	1.4	8.4
			Extract Immature Coarse Grained Volcanic Soil	1.1	6.8
			Coarse Brown Lowland Soil	0.7	4.2
			Brown Forest Soil (IV)	0.7	4.2
			Alpine Soil Debris-Rock Land	0.6	3.5
			Brown Lowland Soil	0.5	2.8
			Brown Forest Soil (III)	0.2	1.4
			Volcano Distillate Immature Soil	0.2	1.3
			High Peat Soil	0.2	1.0
			Brown Forest Soil-Soil Dry Podsolization	0.0	0.2

No.	Dam Catchment	Catchment Area (km ²)	Soil Types	Area (%)	Area (km ²)
11	Gosho	635.17	Rocky Ground	5.6	35.8
			Alpine Debris Soil	0.6	3.9
			Debris Soil	0.2	1.3
			Residual Product of Immature Soil	1.6	10.3
			Thick Layer Black Soil	1.3	8.2
			Andosol	14.3	90.9
			Humid Black Soil	8.5	54.0
			Light Colored Black Soil	2.8	18.0
			Dry Brown Forest Soil	3.4	21.8
			Brown Forest Soil	25.7	163.2
			Moist Brown Forest Soil	2.3	14.4
			Dry Soil Podsolization	4.1	26.3
			Moist Soil Podsolization	20.8	131.9
			Yellow Soil	1.9	12.1
			Fine Grain Gray Lowland Soil	0.4	2.7
			Gray Lowland Soil	1.8	11.5
			Coarse Gray Lowland Soil	0.7	4.4
			Gley Soil Grain Size	1.6	10.2
			Gley Soil	1.0	6.0
			High Peat Soil	1.0	6.6
			Low Peat Soil	0.2	1.5
12	Sagurigawa	61.36	Black Gley Soil	22.8	17.4
			Dry Brown Forest Soil (Yellow)	11.6	8.9
			Dry Brown Forest Soil (red)	1.2	0.9
			Brown Forest Soil	26.6	20.3
			Brown Forest Soil (yellow)	29.6	22.6
			Brown Forest Soil (red)	0.1	0.1
			Gley Soil	8.1	6.1

No.	Dam Catchment	Catchment Area (km ²)	Soil Types	Area (%)	Area (km ²)
13	Ikari	271.20	Rocky Ground	2.8	7.6
			Debris Soil	2.3	6.2
			Thick Layer Black Soil	7.3	19.7
			Andosol	1.7	4.7
			Dry Brown Forest Soil	1.5	4.1
			Brown Forest Soil	68.8	186.6
			Moist Brown Forest Soil	4.0	10.7
			Dry Soil Podsolization	6.3	17.2
			Moist Soil Podsolization	4.2	11.3
			No lakes, rivers	1.2	3.1
14	Kawaji	320.74	Rocky Ground	8.6	27.7
			Debris Soil	4.3	13.8
			Thick Layer Black Soil	5.6	18.0
			Andosol	5.8	18.8
			Dry Brown Forest Soil	4.2	13.6
			Brown Forest Soil	53.7	173.7
			Dry Soil Podsolization	7.9	25.7
			Moist Soil Podsolization	9.7	31.5
			No lakes, rivers	0.1	0.4
15	Kawamata	179.4	Debris Soil	1.0	1.7
			Thick Layer Black Soil	3.4	6.2
			Dry Brown Forest Soil	0.4	0.7
			Brown Forest Soil	20.8	37.4
			Moist Brown Forest Soil	4.4	7.9
			Dry Soil Podsolization	30.3	54.4
			Moist Soil Podsolization	36.3	65.1
			No lakes, rivers	3.4	6.2

No.	Dam Catchment	Catchment Area (km ²)	Soil Types	Area (%)	Area (km ²)
16	Aimata	110.8	Rocky Ground	27.3	30.3
			Alpine Debris Soil	5.0	5.5
			Andosol	7.4	8.2
			Dry Brown Forest Soil	1.3	1.5
			Brown Forest Soil	54.0	59.8
			Moist Soil Podsolization	3.9	4.3
			No lakes, rivers	1.1	1.3
17	Fujiwara	400.2	Rocky Ground	16.4	65.7
			Alpine Debris Soil	2.9	11.4
			Thick Layer Black Soil	0.8	3.2
			Andosol	1.9	7.8
			Dry Brown Forest Soil	1.4	5.7
			Brown Forest Soil	64.6	258.4
			Moist Brown Forest Soil	1.2	4.8
			Dry Soil Podsolization	0.4	1.6
			Moist Soil Podsolization	9.2	36.7
			Coarse Gray Lowland Soil	0.4	1.7
			High Peat Soil	0.0	0.1
			No lakes, rivers	0.8	3.0
18	Kusaki	263.85	Rocky Ground	8.9	23.6
			Debris Soil	11.3	29.8
			Thick Layer Black Soil	0.9	2.2
			Andosol	0.9	2.3
			Dry Brown Forest Soil	2.4	6.4
			Brown Forest Soil	60.4	159.3
			Moist Brown Forest Soil	0.6	1.6
			Dry Soil Podsolization	7.5	19.7
			Moist Soil Podsolization	7.1	18.8
			No lakes, rivers	0.1	0.2

No.	Dam Catchment	Catchment Area (km ²)	Soil Types	Area (%)	Area (km ²)
19	Naramata	95.40	Rocky Ground	21.7	20.7
			Alpine Debris Soil	3.2	3.0
			Thick Layer Black Soil	0.1	0.1
			Dry Brown Forest Soil	1.5	1.4
			Brown Forest Soil	66.1	63.0
			Moist Brown Forest Soil	1.5	1.4
			Dry Soil Podsolization	0.2	0.1
			Moist Soil Podsolization	4.3	4.1
			Coarse Gray Lowland Soil	1.6	1.5
			High Peat Soil	0.1	0.1
20	Shimokubo	323.65	Rocky Ground	0.9	3.0
			Debris Soil	0.5	1.5
			Andosol	2.9	9.5
			Dry Brown Forest Soil	3.9	12.6
			Brown Forest Soil	81.2	262.9
			Moist Brown Forest Soil	7.2	23.1
			Dry Soil Podsolization	1.8	5.7
			Moist Soil Podsolization	1.6	5.2
21	Sonohara	601.06	Rocky Ground	3.4	20.1
			Alpine Debris Soil	0.6	3.8
			Debris Soil	0.8	4.9
			Extract Immature Coarse Grained Volcanic Soil	6.2	37.2
			Thick Layer Black Soil	1.1	6.5
			Andosol	0.7	4.1
			Course Black Soil	12.2	73.2
			Dry Brown Forest Soil	0.6	3.8
			Brown Forest Soil	61.0	366.5
			Moist Brown Forest Soil	1.6	9.4
			Dry Soil Podsolization	1.5	9.0
			Moist Soil Podsolization	8.8	52.9
			Coarse Gray Lowland Soil	1.5	9.3
			No lakes, rivers	0.1	0.4

No.	Dam Catchment	Catchment Area (km ²)	Soil Types	Area (%)	Area (km ²)
22	Yagisawa	165.54	Rocky Ground	17.2	28.5
			Alpine Debris Soil	1.2	2.0
			Brown Forest Soil	60.4	100.0
			Moist Brown Forest Soil	1.0	1.6
			Dry Soil Podsolization	0.1	0.1
			Moist Soil Podsolization	18.3	30.3
			No lakes, rivers	1.8	3.0
23	Futase	170.58	Andosol	0.1	0.2
			Brown Forest Soil	47.1	80.3
			Brown Forest Soil (dark colored)	0.8	1.3
			Moist Brown Forest Soil	20.9	35.7
			Dry Soil Podsolization	30.6	52.2
			Moist Soil Podsolization	0.5	0.8
24	Koshibu	289.57	Rocky Ground	3.9	11.3
			Alpine Debris Soil	1.6	4.6
			Residual Product of Immature Soil	0.7	1.9
			Andosol	0.5	1.6
			Dry Brown Forest Soil	2.1	6.1
			Brown Forest Soil	57.0	164.9
			Moist Brown Forest Soil	18.6	53.8
			Dry Soil Podsolization	11.2	32.4
			Moist Soil Podsolization	3.4	9.8
			Brown Lowland Soil	0.0	0.0
			Gray Lowland Soil	0.5	1.5
			Others (gravel & urban areas)	0.5	1.6

No.	Dam Catchment	Catchment Area (km ²)	Soil Types	Area (%)	Area (km ²)
25	Makio	307.79	Rocky Ground	1.2	3.6
			Thick Layer Black Soil	5.2	15.9
			Andosol	0.5	1.6
			Dry Brown Forest Soil	0.5	1.6
			Brown Forest Soil	29.7	91.3
			Moist Brown Forest Soil	14.7	45.3
			Dry Soil Podsolization	17.4	53.6
			Moist Soil Podsolization	28.3	87.1
			Gray Lowland Soil	1.0	3.2
			Others (gravel & urban areas)	0.5	1.5
			No lakes, rivers	1.0	3.1
26	Miwa	311.03	Rocky Ground	1.4	4.3
			Alpine Debris Soil	1.3	4.0
			Debris Soil	0.4	1.1
			Dry Brown Forest Soil	4.0	12.3
			Brown Forest Soil	47.1	146.4
			Moist Brown Forest Soil	30.5	95.0
			Dry Soil Podsolization	14.0	43.5
			Moist Soil Podsolization	0.9	2.7
			Brown Lowland Soil	0.1	0.2
			Gray Lowland Soil	0.5	1.5

No.	Dam Catchment	Catchment Area (km ²)	Soil Types	Area (%)	Area (km ²)
27	Maruyama	2409	Rocky Ground	1.3	31.3
			Debris Soil	0.3	7.9
			Residual Product of Immature Soil	0.1	2.6
			Coarse Residual Product of Immature Soil	1.4	33.0
			Thick Layer Black Soil	3.2	76.1
			Andosol	4.5	108.6
			Humid Black Soil	1.0	25.1
			Dry Brown Forest Soil	5.6	134.9
			Dry Brown Forest Soil (red)	0.8	19.8
					1121.
			Brown Forest Soil	46.6	9
			Brown Forest Soil (dark colored)	0.0	0.0
			Moist Brown Forest Soil	9.3	223.1
			Dry Soil Podsolization	12.8	307.4
			Moist Soil Podsolization	6.8	163.6
			Red Soil	0.2	4.1
			Yellow Soil	1.4	33.0
			Fine Grain Gray Lowland Soil	2.4	57.6
			Gray Lowland Soil	0.8	20.2
			Coarse Gray Lowland Soil	0.1	2.6
			Gley Soil Grain Size	0.8	20.2
			Others (gravel & urban areas)	0.5	12.5
			No lakes, rivers	0.1	3.4
28	Yokoyama	470.71	Debris Soil	0.9	4.4
			Humid Black Soil	0.3	1.6
			Dry Brown Forest Soil	13.1	61.7
			Brown Forest Soil	70.0	329.5
			Moist Brown Forest Soil	2.7	12.7
			Dry Soil Podsolization	7.3	34.3
			Moist Soil Podsolization	4.3	20.3
			Gray Lowland Soil	0.7	3.2
			Gley Soil Coarse	0.7	3.2

No.	Dam Catchment	Catchment Area (km ²)	Soil Types	Area (%)	Area (km ²)
29	Sintoyone	111.44	Dry Brown Forest Soil	42.9	47.8
			Brown Forest Soil	57.1	63.6
30	Yahagi	504.62	Dune Immature Soil	0.3	1.5
			Thick Layer Black Soil	0.6	3.0
			Andosol	2.0	10.1
			Dry Brown Forest Soil	27.5	138.6
			Brown Forest Soil	58.2	293.8
			Moist Brown Forest Soil	2.8	14.2
			Dry Soil Podsolization	0.8	3.8
			Moist Soil Podsolization	1.6	7.9
			Yellow Soil	0.3	1.6
			Brown Lowland Soil	0.3	1.6
			Coarse Brown Lowland Soil	1.0	5.2
			Gray Lowland Soil	4.6	23.2
31	Hitokura	115.1	Residual Product of Immature Soil	13.1	15.0
			Dry Brown Forest Soil	7.6	8.8
			Dry Brown Forest Soil (Yellow)	3.8	4.3
			Dry Brown Forest Soil (red)	68.5	78.9
			Brown Lowland Soil	1.3	1.5
			Coarse Brown Lowland Soil	1.4	1.6
			Fine Grain Gray Lowland Soil	2.9	3.3
			Gray Lowland Soil	1.4	1.6
32	Ishitegawa	72.6	Dry Brown Forest Soil	53.6	38.9
			Brown Forest Soil	46.4	33.7

No.	Dam Catchment	Catchment Area (km ²)	Soil Types	Area (%)	Area (km ²)
33	Nomura	168.00	Dry Brown Forest Soil	40.0	67.1
			Dry Brown Forest Soil (Yellow)	13.6	22.8
			Brown Forest Soil	29.3	49.1
			Yellow Soil	1.0	1.6
			Coarse Brown Lowland Soil	1.9	3.2
			Fine Grain Gray Lowland Soil	7.6	12.8
			Gray Lowland Soil	5.8	9.7
			Gley Soil Grain Size	1.0	1.6
34	Kyuuragi	33.7	Dry Brown Forest Soil	1.0	0.3
			Dry Brown Forest Soil (Yellow)	17.0	5.7
			Brown Forest Soil (yellow)	55.5	18.7
			Brown Forest Soil (red)	3.0	1.0
			Moist Brown Forest Soil	10.5	3.5
			Yellow Soil	3.5	1.2
			Coarse Gray Lowland Soil	9.6	3.2

Appendix 3. Summary of Terrain Types

No.	Dam Catchment	Catchment Area (km ²)	Terrain Types	Area (%)	Area (km ²)
1	Houheikyou	136.1	Rugged Mountain	33.4	45.5
			Small Relief Mountain	66.6	90.6
2	Iwaonai	341.58	Large Undulating Mountain	6.3	21.5
			Rugged Mountain	39.1	133.5
			Small Relief Mountain	47.8	163.3
			Flat Alluvial Valley	3.0	10.1
			Gravel Terrace (Medium Old)	0.8	2.8
			Mesa (Very Old)	1.8	6.0
			Mesa (Medium Old)	1.3	4.3
3	Izarigawa	113.25	Rugged Mountain	30.7	34.8
			Small Relief Mountain	19.5	22.1
			Large Undulating Volcanic	0.6	0.7
			Large Undulating Hill	49.1	55.7
4	Jyouzankei	103.59	Rugged Mountain	60.2	62.3
			Small Relief Mountain	39.9	41.3
5	Kanayama	410.81	Large Undulating Mountain	11.4	47.0
			Rugged Mountain	8.3	34.2
			Small Relief Mountain	12.7	52.3
			Piedmond Area	1.6	6.5
			Large Undulating Volcanic	3.9	16.1
			Medium Undulating Volcanic	7.6	31.2
			Small Undulating Volcanic	3.7	15.2
			Volcanic Foot Hill	3.5	14.3
			Large Undulating Hill	0.6	2.4
			Small Undulating Hill	5.7	23.5

No.	Dam Catchment	Catchment Area (km ²)	Terrain Types	Area (%)	Area (km ²)
6	Nibutani	1155.45	Flat Alluvial Valley	13.7	56.2
			Loam Terrace (Very old)	0.2	0.9
			Loam Terrace (Old)	0.1	0.3
			Gravel Terrace (Very Old)	0.1	0.6
			Gravel Terrace (Medium Old)	1.9	7.7
			Gravel Terrace (Old)	3.5	14.3
			Mesa (Very Old)	21.0	86.3
			Mesa (Medium Old)	0.4	1.5
			Mesa (Old)	0.1	0.3
			Large Undulating Mountain	29.1	353.9
			Rugged Mountain	32.6	395.6
			Small Relief Mountain	28.1	341.3
			Piedmond Area	0.4	4.5
			Large Undulating Hill	1.5	18.0
			Flat Alluvial Valley	2.4	29.5
			Gravel Terrace (Medium Old)	0.8	9.4
			Gravel Terrace (Old)	5.2	62.8
7	Pirika	114.44	Rugged Mountain	17.9	20.5
			Small Relief Mountain	82.1	94.0
8	Satsunaigawa	116.63	Large Undulating Mountain	100.0	116.6
9	Taisetsu	289.26	Large Undulating Mountain	9.4	27.1
			Rugged Mountain	17.4	50.2
			Small Relief Mountain	19.3	55.9
			Large Undulating Volcanic	14.0	40.4
			Medium Undulating Volcanic	22.5	65.1
			Flat Alluvial Valley	3.9	11.3
			Gravel Terrace (Very Old)	3.4	9.8
			Gravel Terrace (Medium Old)	9.2	26.6
			Gravel Terrace (Old)	1.0	2.8

No.	Dam Catchment	Catchment Area (km ²)	Terrain Types	Area (%)	Area (km ²)
10	Tokachi	598.24	Large Undulating Mountain	3.9	23.3
			Rugged Mountain	61.8	369.6
			Small Relief Mountain	23.4	139.8
			Large Undulating Volcanic	7.2	42.8
			Medium Undulating Volcanic	3.8	22.7
			Mesa (Very Old)	0.0	0.1
11	Gosho	635.17	Large Undulating Mountain	10.3	65.7
			Rugged Mountain	30.3	192.6
			Small Relief Mountain	5.8	37.1
			Large Undulating Volcanic	4.0	25.4
			Medium Undulating Volcanic	11.6	73.6
			Small Undulating Volcanic	13.0	82.8
			Volcanic Foot Hill	4.3	27.2
			Small Undulating Hill	8.2	52.3
			Volcanic Hills	5.7	36.1
			Gravel Terrace	5.5	35.2
			Flat Alluvial Valley	1.1	7.2
12	Sagurigawa	61.36	Small Relief Mountain	24.2	18.5
			Small Undulating Volcanic	9.0	6.9
			Large Undulating Hill	45.6	34.7
			Small Undulating Hill	1.1	0.8
			Messa terrace	7.2	5.5
			Flat Alluvial Valley	13.0	9.9
13	Ikari	271.2	Large Undulating Mountain	51.4	139.4
			Rugged Mountain	43.2	117.1
			Small Relief Mountain	4.4	11.9
			Medium Undulating Volcanic	0.2	0.5
			Small Undulating Volcanic	0.3	0.9
			Flat Alluvial Valley	0.5	1.5

No.	Dam Catchment	Catchment Area (km ²)	Terrain Types	Area (%)	Area (km ²)
14	Kawaji	320.74	Large Undulating Mountain	12.6	40.6
			Rugged Mountain	57.1	184.6
			Large Undulating Volcanic	21.6	69.6
			Medium Undulating Volcanic	8.6	27.7
			Lake	0.2	0.6
15	Kawamata	179.4	Large Undulating Mountain	30.4	54.6
			Rugged Mountain	64.2	115.1
			Large Undulating Volcanic	2.7	4.8
			Medium Undulating Volcanic	1.1	2.0
			Lake	1.6	2.9
16	Aimata	110.8	Large Undulating Mountain	49.6	55.0
			Rugged Mountain	40.7	45.1
			Small Relief Mountain	1.5	1.7
			Gravel Terrace	1.5	1.6
			Delta Lowland	5.5	6.1
			Lake	1.1	1.3
17	Fujiwara	400.2	Large Undulating Mountain	55.6	222.7
			Rugged Mountain	22.1	88.4
			Small Relief Mountain	1.6	6.3
			Piedmond Area	0.4	1.5
			Large Undulating Volcanic	13.5	54.1
			Medium Undulating Volcanic	5.4	21.5
			Small Undulating Volcanic	0.7	2.6
			Lake	0.8	3.1
18	Kusaki	263.85	Large Undulating Mountain	55.0	145.1
			Rugged Mountain	40.3	106.4
			Small Relief Mountain	2.3	6.0
			Flat Alluvial Valley	1.7	4.5
			Delta Lowland	0.6	1.6
			Lake	0.1	0.2

No.	Dam Catchment	Catchment Area (km ²)	Terrain Types	Area (%)	Area (km ²)
19	Naramata	95.4	Large Undulating Mountain	27.6	26.3
			Rugged Mountain	22.9	21.8
			Small Relief Mountain	1.0	1.0
			Large Undulating Volcanic	35.0	33.4
			Medium Undulating Volcanic	11.0	10.5
			Small Undulating Volcanic	2.5	2.3
20	Shimokubo	323.65	Large Undulating Mountain	50.2	162.6
			Rugged Mountain	48.7	157.5
			Small Relief Mountain	1.1	3.6
21	Sonohara	601.06	Large Undulating Mountain	39.7	238.4
			Rugged Mountain	28.7	172.3
			Small Relief Mountain	10.8	65.1
			Large Undulating Volcanic	6.0	35.9
			Medium Undulating Volcanic	8.4	50.5
			Small Undulating Volcanic	2.0	12.2
			Loam Terrace, Hill section	1.8	10.8
			Gravel Terrace	0.3	1.6
			Delta Lowland	2.3	13.9
			Lake	0.1	0.4
22	Yagisawa	165.54	Large Undulating Mountain	78.0	129.2
			Rugged Mountain	20.2	33.4
			Lake	1.8	3.0
23	Futase	170.58	Large Undulating Mountain	93.8	160.0
			Rugged Mountain	6.2	10.6
24	Koshibu	289.57	Large Undulating Mountain	86.0	249.0
			Rugged Mountain	9.0	25.9
			Small Relief Mountain	4.0	11.5
			Flat Alluvial Valley	1.1	3.2

No.	Dam Catchment	Catchment Area (km ²)	Terrain Types	Area (%)	Area (km ²)
25	Makio	307.79	Large Undulating Mountain	14.8	45.7
			Rugged Mountain	29.9	92.1
			Small Relief Mountain	24.0	73.8
			Large Undulating Volcanic	14.7	45.2
			Medium Undulating Volcanic	10.6	32.6
			Small Undulating Volcanic	3.6	11.0
			Volcanic Foot Hill	0.9	2.7
			Gravel Terrace	0.5	1.6
			Lake	1.0	3.1
26	Miwa	311.03	Large Undulating Mountain	85.2	265.0
			Rugged Mountain	12.4	38.5
			Small Relief Mountain	2.4	7.5
27	Maruyama	2409.00	Large Undulating Mountain	34.7	835.9
			Rugged Mountain	23.5	566.1
			Small Relief Mountain	19.2	463.0
			Piedmond Area	3.3	78.5
			Large Undulating Volcanic	2.9	69.9
			Medium Undulating Volcanic	1.9	46.7
			Small Undulating Volcanic	1.2	27.9
			Volcanic Foot Hill	0.2	5.8
			Large Undulating Hill	2.9	68.9
			Small Undulating Hill	4.9	116.8
			Gravel Terrace, Hill section	0.1	1.4
			Gravel Terrace	2.3	56.1
			Flat Alluvial Valley	2.9	68.9
			Lake	0.1	2.9
28	Yokoyama	470.71	Large Undulating Mountain	51.0	239.8
			Rugged Mountain	48.4	227.7
			Flat Alluvial Valley	0.7	3.2

No.	Dam Catchment	Catchment Area (km ²)	Terrain Types	Area (%)	Area (km ²)
29	Sintoyone	111.44	Large Undulating Mountain	7.6	8.5
			Rugged Mountain	30.6	34.1
			Small Relief Mountain	61.2	68.2
			Flat Alluvial Valley	0.6	0.6
30	Yahagi	504.62	Large Undulating Mountain	2.6	13.0
			Rugged Mountain	47.7	240.6
			Small Relief Mountain	48.2	243.0
			Flat Alluvial Valley	1.6	7.9
31	Hitokura	115.10	Rugged Mountain	27.2	31.3
			Small Relief Mountain	62.8	72.2
			Large Undulating Hill	0.9	1.1
			Small Undulating Hill	0.7	0.8
			Flat Alluvial Valley	8.4	9.7
32	Ishitegawa	72.60	Large Undulating Mountain	12.4	9.0
			Rugged Mountain	41.1	29.9
			Small Relief Mountain	45.6	33.1
			Small Undulating Hill	1.0	0.7
33	Nomura	168.00	Rugged Mountain	36.8	61.8
			Small Relief Mountain	47.9	80.5
			Flat Alluvial Valley	15.3	25.8
34	Kyuuragi	33.7	Large Undulating Mountain	2.4	0.8
			Rugged Mountain	30.2	10.2
			Small Relief Mountain	67.5	22.7

RESUME

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LIST OF PUBLICATIONS

Peer-Reviewed Papers

1. Intan Supraba and Tomohito J. Yamada, “Catchment Storage Estimation Based on Total Rainfall-Total Loss Rainfall Relationship for 47 Catchments in Japan”, *Journal of Japan Society of Civil Engineers, Ser.B1 (Hydraulic Engineering)*, Vol. 70, No. 4, I_169-I_174, 2014.
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3. Intan Supraba and Tomohito J. Yamada, “Uncertainty of Peak Runoff Height Associated With Uncertainties of Water Holding Capacity and Rainfall Pattern”, *Journal of Global Environment Engineering, Japan Society of Civil Engineers*, 2015.

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4. Intan Supraba, Takenori Kouno, Dwi Prabowo Yuga Suseno, Yadu Pokhrel, Tomohito J. Yamada, “Development of MATSIRO-Land Surface Model- for Higher Resolution in Local Scale (Ishikari Basin Area in Hokkaido, Japan), in The 6th Asia Pacific Association of Hydrology and Water Resources (APHW), 19th-21st August 2013, Korea University, Seoul, Korea.
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9. Intan Supraba and Tomohito J. Yamada, “Uncertainty of Peak Runoff Height Associated With Uncertainties of Water Holding Capacity and Rainfall Pattern”, in Japan Geoscience Union (JpGU) Meeting 2015, 24th-28th May 2015, Makuhari Messe, Japan.