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# Ecohydrology and Biogeochemistry in a Temperate Forest Catchment

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## 1. Introduction

The assimilated carbon stored in terrestrial ecosystems is exported with water movement in both organic and inorganic forms, which are defined as particulate organic carbon (POC), dissolved organic carbon (DOC), and dissolved inorganic carbon (DIC). The transport of terrestrial carbon into streams, rivers and eventually the oceans is an important link in the global carbon cycle (Ludwig et al., 1996; Warnken and Santschi, 2004). The Committee on Flux of Carbon to the Ocean estimated that of the organic carbon entering rivers globally, around 50% is transported to the ocean, 25% is oxidized within the system and 25% stored as POC in the system as sediment (Hope et al. 1994). As compared to the terrestrial carbon sinks (1.9 Gt-C/yr; Prentice et al., 2001), the organic carbon transport from terrestrial ecosystems to oceans is 0.4–0.9 Gt-C/yr (Meybeck, 1982; Hope et al., 1994; Prentice et al., 2001), representing a substantial component of the ecosystem carbon balance.

The water and carbon cycles in forest catchments are important elements for understanding the impact of global environmental changes on terrestrial ecosystems. Various theories have been suggested to better understand water discharge (Horton, 1933; Betson, 1964; Kirkby, 1978; Anderson and Burt, 1991; Kim et al., 2003) and its effect on carbon efflux processes from forest catchments (McGlynn and McDonnell, 2003; Kawasaki et al., 2005; Schulze, 2006; Kim et al., 2007b; Kim et al., 2010). Most of the results indicated that the hydrological flowpaths are important in carbon dynamics within the forest catchments.

Data from major results show export of organic carbon to be highly correlated with annual river discharge and watershed size (Table 1; Fig. 1). Hydrological processes strongly affect organic carbon discharge from terrestrial ecosystems, especially in monsoon climate zone of East Asia, and 60-80% of annual organic carbon export to the ocean during summer rainy season (Tao, 1998; Liu et al., 2003; Kawasaki et al., 2005; Zhang et al., 2009; Kim et al., 2010).

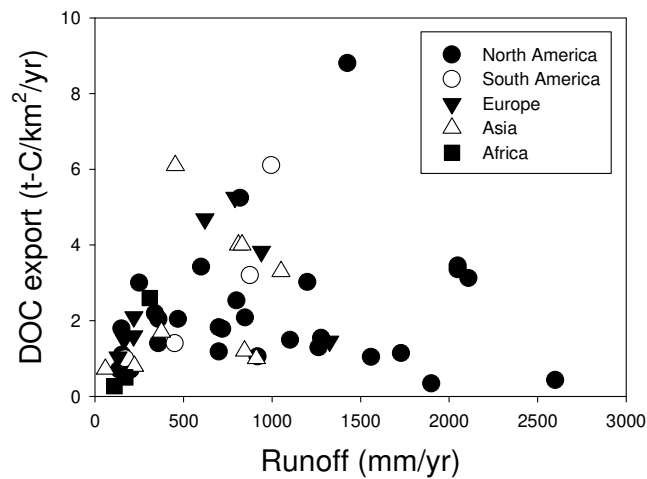


Fig. 1. Annual DOC export by rivers from watersheds (modified from Ittekkot and Laane (1991), Hope et al. (1994), and Table 1)

Ecosystem type, location	Annual precipitation (mm)	Annual runoff (mm)	Watershed size (km <sup>2</sup> )	Loss of organic carbon		Reference
				DOC (t-C/ha/yr)	POC (t-C/ha/yr)	
Huanghe, Semi-arid area, Central China	-	59	745,000	0.007	-	Gan et al. (1983)
Yichun, Humid temperate area, North-eastern China	500-650	-	2500	0.03	-	Tao (1998)
Luodingjiang River, Subtropical mountainous, Southern China	1534	844	3164	0.012	0.011	Zhang et al. (2009)
Guandaushi, Subtropical forest, Taiwan	2300-2700	-	0.47	0.025	-	Liu et al. (2003)
Tomakomai, Cool temperate mixed forest, Northern Japan	1200	-	9.4	0.0052	0.0076	Shibata et al. (2005)
Kiryu, Temperate conifer, Central Japan	1645	911	0.006	0.01	-	Kawasaki et al. (2005)
Gwangneung, Temperate deciduous, Central Korea	1332	809	0.22	0.04	0.05	Kim et al. (2010)
Han River, Temperate area, Central Korea	1244	-	26,018	0.04	0.02	Kim et al. (2007a)

Table 1. Export of organic carbon East Asian watersheds

Forests are the major terrestrial biome, in which soils and vegetation are the primary sources of DOC and POC in streamwater. Within the forest soil profile, concentrations of DOC typically are highest in the interstitial waters of the organic-rich upper soil horizons (McDowell and Likens, 1988; Richter et al., 1994; Dosskey and Bertsch, 1997). Both column experiments and field observations have indicated that significant transport of DOC occurs by preferential flow, given that the state of adsorption equilibrium cannot be reached, owing to the reduction of the contact time between DOC and the soil surface (Jardine et al., 1989; Hagedorn et al., 1999). Understanding the flow paths of DOC discharge from forested catchments to streams is important because DOC provides a source of energy to microorganisms in water systems (Stewart and Wetzel, 1982) and carbon fixation in the soil (Neff and Asner 2001; Kawasaki et al., 2005).

Since the 1970s 2.3 million hectares of land have been planted with coniferous species such as *Pinus Koraiensis*, *Abies holophylla* and *Larix leptolepis* in South Korea. Because coniferous forests lose and consume water resources much more than deciduous forests due to higher leaf area index (LAI) and year-round transpiration, these planted coniferous forests may deteriorate the physical properties of the topsoil owing to the water repellence of the soil surface and decrease the availability of water resources by high evapotranspiration rates. To conserve soil and water resources, densely planted coniferous forests must be managed using silvicultural techniques (e.g. pruning, thinning) that could influence water quantity and quality. In South Korea, various studies have shown that forest management practices in coniferous forests decreased the amount of interception loss and increased discharges during the dry season. Clear cutting resulted in catastrophic augmentation of runoff and soil loss. The water quality of stream headwaters improved after thinning and pruning because these techniques tended to ameliorate soil physical properties and increase soil ion exchange capacity.

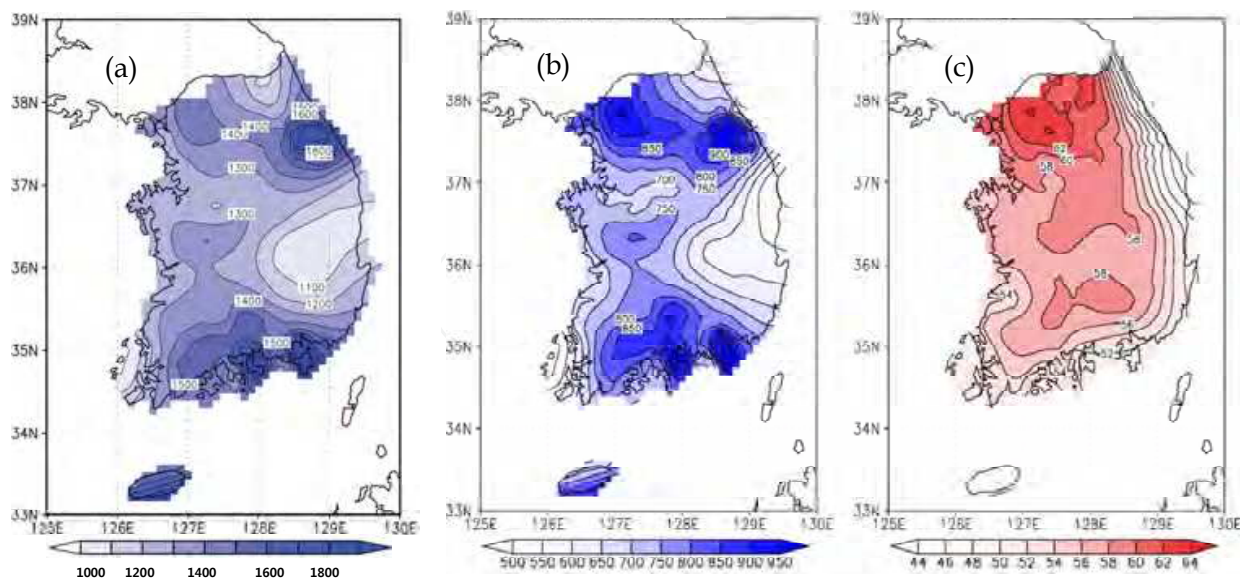


Fig. 2. Accumulated climatological precipitation (mm) at 60 stations for (a) annual total precipitation, (b) summer (from June to August) in South Korea. (c) Percentage (%) of accumulated climatological precipitation with respect to annual total precipitation summer. (Seo et al., 2011)

Hydrological circumstances in South Korea are unfavourable to manage water resources. Temporal and spatial variations of rainfall are very large (Fig. 2). Annual rainfall ranges from 754 to 1,683 mm. In South Korea, more than 50% of the annual precipitation falls in the summer monsoon season (Fig. 2(b, c)), which quickly discharges to the ocean due to the steep slopes and short river lengths (<500 km). Therefore, the water regime in the catchment undergoes drastic changes with recurring wet and dry seasons, which makes it difficult to interpret and predict hydrological processes and subsequently their effect on nutrients cycling (Kim et al., 2009).

The amount of water storage capacity in a forest stand increases with the forest aging, when a forest stand grows, the amount of litter falls and roots also increases. The mineral soils and humus materials tend to aggregate into their structure. The aggregation may change the distribution of pore sizes and often increase the total porosity of soil. Forests have been called 'Green Dam' or 'Reservoir' because of its function of controlling the flood and drought through a litter layer and topsoil like a sponge filter. Net infiltration rate of a well-developed forest soil has been estimated 76 mm/hr in comparison with 8mm/hr of a bare land (Brooks et. al. 1991). Generally, infiltration capacity of soil in a deciduous forest is higher than that in a coniferous forest because the litter fall of the former is easily decomposed and incorporated with mineral particles compared to that of the latter. Most of stream water in South Korea comes from mountain headwaters as mountains occupy 65% of the total area. Stream in forested headwaters yields clean water. Forest soils hold the water like sponge, which is 3.3 times more than the soil in a bare land. Water holding capacity of forest soil in Korea is estimated about 18 billion tons, as shown in Table 2 (Ministry of Science and Technology, 1992).

Bed rock	Igneous	Metanorphic	Basalt	Sedimentary		Lime	Total
				I	II		
Maximum storage capacity(%)	A	34.3	40.1	32.4	36.1	33.9	39.8
	B	39.5	44.4	35.1	39.0	35.5	41.1
Total storage (0.1 billion tons)	A	15.2	20.0	5.0	4.0	1.5	2.1
	B	36.5	60.3	8.5	14.2	4.6	7.9
Sub-total		51.7	80.3	13.5	18.2	6.1	179.8

Table 2. Water holding capacity of forest soils depending on bed rock in South Korea (Ministry of Science and Technology, 1992)

Despite decades of dedicated scientific efforts on these fundamental questions, it is still difficult to find a robust interpretation even for some basic hydrological processes such as discharge and runoff. The up to date results showed that the geophysical and meteorological conditions greatly affect the hydrological processes (Hooper et al., 1990; Elsenber et al., 1995; Katsuyama et al., 2001; McGlynn and McDonnell, 2003; Kim et al., 2010).

In this chapter, we have implemented a comprehensive ecohydrological measurement system at the temperate forest catchment in South Korea. Most importantly, high quality long-term data of hydrological and meteorological conditions have been collected, which may be also important in monitoring global environmental changes and their effects. The study was also designed based on a nested watershed concept (smaller catchments are

nested in successively larger catchments) to investigate how catchment processes change as scale varies. In this chapter, we introduce the concepts and techniques that were implemented to investigate the movement of water and carbon in a forest catchment. We also briefly discuss preliminary results and their implications for the interactions between hydrological and biogeochemical processes in a temperate forest catchment.

## 2. Hydrological cycle of forested catchments in South Korea

### 2.1 Interception loss and evapotranspiration

The differences in the amount and process of interception loss and evapotranspiration depend on the factors of the forest structure and local climate. The forest structure includes the forest type, age and density. Generally, coniferous forests intercept rain and snowfall more than deciduous because the former has higher LAI and longer leaf-period than the latter.

The first research on interception loss by tree canopy and stem in Korea had conducted in 1935 for determining the total and net precipitation at the forest stand in Korea Forest Research Institute (Kim and Jo, 1937). The experiment was conducted during 23 months on the natural 50-year-old red pine (*Pinus densiflora*) with the tree height of 12 m and DBH of 18 cm. It showed that the annual total and net rainfall were 1,194.2 mm and 1,066.3 mm, respectively. The percentage of interception loss from the total rainfall varied from less than 10% in the season to more than 26% in the growing dormant season.

To clarify the effects of forest types on interception, three types of forest were chosen in Gwangnung experiment station during the period of 1982 to 1988, namely natural matured-deciduous, planted young-coniferous and rehabilitated mixed forest. The results of the research are shown in Table 3 (Lee et al. 1989). Among the three forest types, the planted young-coniferous forest showed the most interception loss of 32.6%, compared to 29.1% in natural matured-deciduous and 18.5% in rehabilitated mixed forest. Even though the naturally matured deciduous has the largest forest structure of 80 years old, its amount of interception loss resulted in less percentage compared with the planted young-coniferous for the cause mentioned above.

Forest type	Precipitation (mm)	Intensity (mm/hr)	Throughfall (mm)	Steamflow (mm)	Interception (mm)	Interception (%)
Mixed	1733.2	6.7	1312.6 (75.8)	99.2 (5.7)	321.4	18.5
Coniferous	1477.3	6.0	945.2 (64.0)	50.4 (3.4)	481.5	32.6
Deciduous	1172.4	6.2	758.7 (64.7)	72.2 (6.2)	341.5	29.1

( ) means % for precipitation

Table 3. The amount of interception loss by three forest types in South Korea

In other results, the young coniferous forest of 26-year-old *Pinus rigitaeda* and deciduous forest of 16-year-old *Quercus mongolica* intercepted 17.4 and 13.9% of the total rainfall,

respectively, during the period of July 1986 to September 1987 in Seoul National University's Gwanak arboretum (Kim and Woo 1988a, b).

It is difficult to measure the exact amount of interception loss due to the large variations of forests and climate factors. Several forest hydrologists have tried to predict the amount by using an interception model. There are three kinds of interception model for the estimation of the processes and amount of interception; the dynamic, analytical and regression-methods. The dynamic-interception model was developed using the forest stand structure and Penman-Monteith model to predict the amount of evaporation under saturation condition (Kim and Woo 1997).

Another loss component of hydrological cycle in forested catchment is evapotranspiration from tree canopy during a period of no rainfall. The amount of evapotranspiration can be estimated by using the water budget method in a short term or a calculating method like penman or Thornthwaite method. The amount of evapotranspiration estimated by Thornthwaite method in three forest types is shown in Table 4 (Kim 1987).

(unit: ton/day/ha)

Type	Month											
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Yangju <sup>a</sup>	0.0	0.0	2.7	15.7	31.0	38.4	48.5	47.7	31.2	15.2	3.6	0.0
Mixed <sup>b</sup>	3.5	4.1	8.7	11.5	16.3	18.4	25.7	26.1	20.6	10.5	10.7	7.3
Gwangnung <sup>a</sup>	0.0	0.0	3.2	17.8	31.4	40.3	50.6	49.9	32.4	17.0	4.0	0.2
Coniferous <sup>b</sup>	6.3	6.6	15.7	14.0	33.3	34.0	44.4	56.4	40.3	17.8	12.0	8.2
Deciduous <sup>b</sup>	6.0	4.2	11.5	10.8	20.9	35.7	17.9	29.3	24.6	17.8	10.7	8.2

<sup>a</sup> means the amount of evapotranspiration by Thornthwaite method.

<sup>b</sup> means the amount of evapotranspiration by short-term water budget method.

Table 4. Monthly evapotranspiration determined by means of Thornthwaite and short-term water budget methods.

## 2.2 Discharge and soil loss variations depending on land cover

In the 1960s and 70s, the main theme on forest hydrology was to evaluate soil and water conservation at the different land cover types. Because in those periods, most of the lands in South Korea were completely devastated all over the country, development of techniques on the erosion control was urgent, especially in the fields (Fig. 3).

Lee et al. (1967) clarified that land cover type influence discharge in the small plot. They concluded that the coniferous plot produces the least discharge (26%) while bare land produces most (76%). They also found that the discharge in the bare land plot started at the rainfall of 10 mm and increased radically at the rainfall more than 80 mm, whereas in the coniferous plot started at the rainfall of 30 mm.

Kim (1987) estimated effects of floods, direct flow reduction and long-term yields on forest by utilizing the measured rainfall-runoff data from the three above-mentioned experimental catchments. He found that the flood peak discharge of the young coniferous and mature deciduous stands were 49% and 36% of the devastated-mixed stand respectively. Also, the

direct flow dropped to 53% in the young coniferous stand and to 55% in the mature deciduous stand, compared to that of the devastated-mixed catchment.

Lee et al. (1989) analyzed the runoff rate and soil at the the natural deciduous, the planted coniferous and the rehabilitated mixed forests, using the data from 1980 to 1988. The runoff rates of the three forest types in an order above were 61.9%, 48.5% and 71.3%, respectively. They concluded that the natural deciduous forest mitigated the peak of flow during the rainy season while it discharged more of low-flow during the dry season, in comparison with the rehabilitated mixed forest. The amount of soil loss during the rainy season was the highest in the rehabilitated mixed forest (2.2 ton/ha/yr) and the least in the deciduous one (0.7 ton/ha/yr).



Fig. 3. Before and after the construction of a hillside planting work in South Korea

Several techniques for analyzing the discharge components include surface runoff, interflow and groundwater, for different forest types in a long-term. The recession coefficient represents the rate of runoff that is released from a soil and streamside. If the recession coefficient of an independent event in the hydrograph changes statistically with the lapse of time, the hydrological characteristics of the forested catchment would be changed. Korea Forest Research Institute (1998) studied the hydrological variation of discharge, soil loss and recession coefficient in three small, forested catchments, using a long-term hydrological data from 1983 to 1992. This study included the naturally matured deciduous, planted coniferous and erosion-controlled mixed forest. The amount of discharge and soil loss varied with the rainfall and forest type. Fig. 4 (up left) shows the variation of the recession coefficient of surface runoff ( $\alpha_1$ ) for 10 years.  $\alpha_1$  gradually decreases in the coniferous forest while it does not show the tendency in the others. This may be caused by the change of the forest structure in the coniferous forest after the planting. The amount of the initial loss by the interception and transpiration has been greatly increased since 1976 as the coniferous trees grow. However, the forest structures in others have not much changed since 1983.

The recession coefficient of interflow ( $\alpha_2$ ) decreased in the coniferous and mixed forests with time Fig. 4 (up right). This can be interpreted by an increase in the soil storage capacity after the planting and erosion control work. As the amount of evapotranspiration increases, the storage opportunity of rainfall in the soil improves. Increment of the storage capacity may result in delaying the releasing time of interflow from the soil.



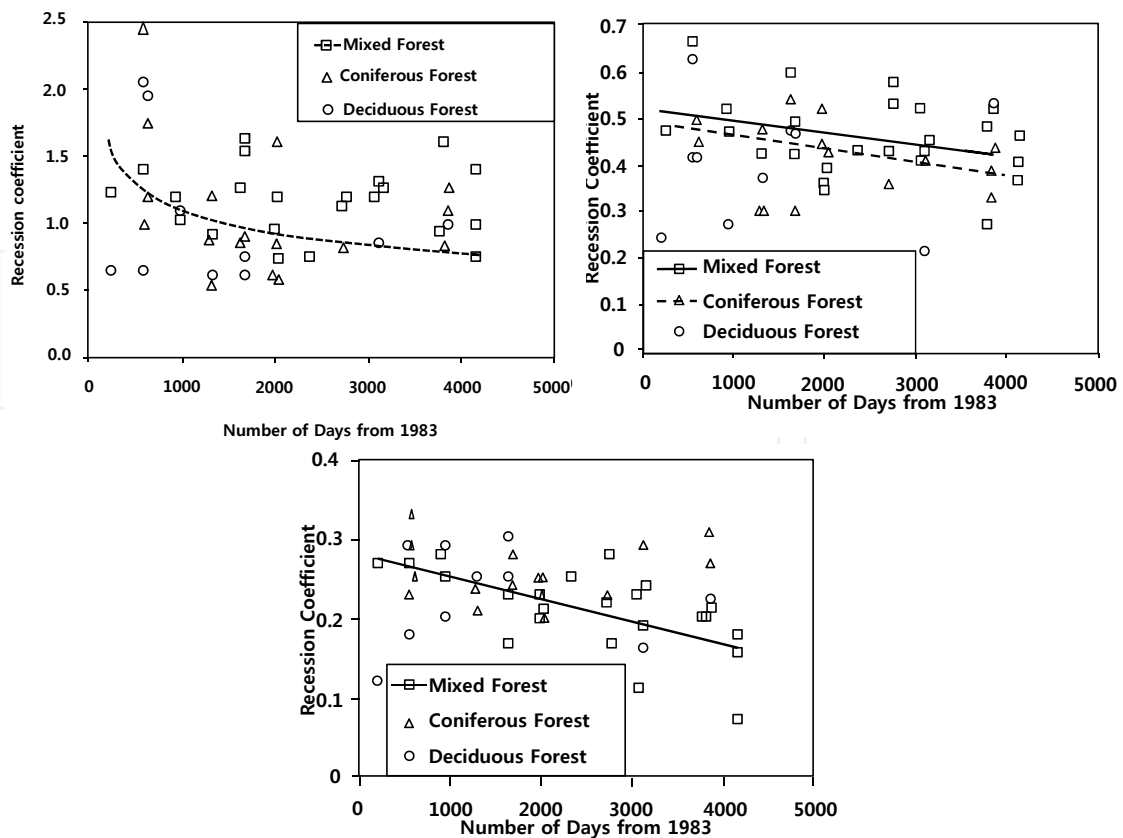


Fig. 4. Variation of the recession coefficient of surface runoff (up left), interflow (up right), and groundwater (down) for 10 years

In the case of the recession coefficient of groundwater ( $\alpha_3$ ), only mixed forest showed a gradual reduction for 10 years (Fig. 4. down). The mixed forest has been a devastated land until erosion control work had finished in 1974. After the work, the soil layer rapidly formed and the soil's physical properties improved.

### 2.3 Residence time of water in a forest catchment

Various radioactive tracers have provided valuable information regarding hydrological processes, such as mean residence time of water, flowpaths during storm events, groundwater movement, and biogeochemical reactions occurring along the flowpaths (Michel and Naftz, 1995; Shanley et al., 1998; Sueker et al., 1999). For example,  $^3\text{H}$  and  $^{14}\text{C}$  have been widely used for determination of time scale of hydrological processes (Matsutani et al., 1993). However, these tracers are inadequate for studying hydrological processes in small and headwater catchments with expected time scales of a year or less because of their long half lives (decades to thousands of years). In this study, we will introduce a short-lived cosmogenic radioactive isotope of  $^{35}\text{S}$  (half life = 87 days) for measuring the mean residence time of water in the Gwangneung catchment.

The measured activity of  $^{35}\text{S}$  in water can be expressed as an equation:

$$C = C_0 e^{-\lambda t} \quad (1)$$

where  $C_0$  is the initial  $^{35}\text{S}$  activity,  $\lambda$  is the decay constant (0.0079655),  $t$  is the number of days from the start of decay, and  $C$  is the measured  $^{35}\text{S}$  activity. The  $^{35}\text{S}$  activity in water provided information of the residence time of atmospherically deposited sulfate. Biogeochemical reactions such as adsorption/desorption in soil and groundwater are also important in affecting the calculated residence time of water in a forested catchment. Assuming a conservative response of sulfate in streamwater, the mean residence time of water was < 40 days during the summer monsoon period in the natural deciduous forest catchment. However, the mean residence time of water increased to around 100 days in the dry season with increasing contribution of the base flow to the stream water (Fig. 5). These results demonstrate that  $^{35}\text{S}$  is useful in estimating the age of water exiting a small catchment where the time scales of hydrologic processes are on the order of 1 year or less.

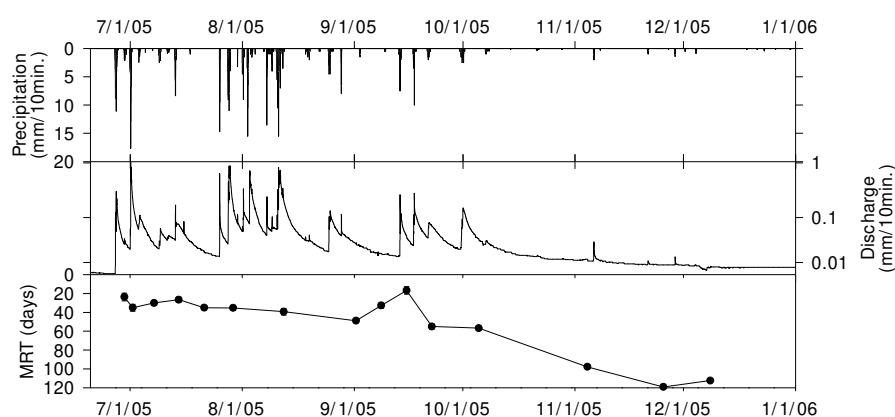


Fig. 5. Temporal variation in mean residence time (MRT) calculated from  $^{35}\text{S}$  based method along with the precipitation and stream discharge (Kim et al., 2009)

From this MRT estimate, the existence of substantial, long-term subsurface water storage is not supported in the studied catchment. The assumed rapid turnover of water in the catchment indicates that the hydrological conditions will respond to the change in precipitation directly and immediately. Therefore, surplus (flooding) and shortage (drought) of water supply may alternate at a relatively short time scale (even within a year) depending on the seasonal distribution of precipitation. A secure water resource planning in catchments of this type will require a reliable prediction and efficient management of precipitation and surface water bodies (Kim et al., 2009).

#### 2.4 Flow paths of water during storm events

The identification of flow paths in forested catchments has been elusive because of difficulties in measuring subsurface flow. Forested catchments are spatially complex and subsurface flow is invisible. Hence, one can only infer the movement and mixing of water from the natural tracer elements that the water carries (Pinder and Jones, 1969). Using various tracers, the end-member mixing analysis (EMMA) has been used to elucidate flow paths and hydrological processes in several catchments (e.g. Hooper et al., 1990; Christophersen et al., 1990; Elsenbeer et al., 1995; Katsuyama et al., 2001). Numerous conceptual models have adopted the flow path dynamics proposed by Anderson et al. (1997), i.e., both pre-event soil water and bedrock groundwater contribute to the formation

of a saturated zone in the area adjacent to the stream (e.g., McGlynn et al., 1999; Bowden et al., 2001; Uchida et al., 2002).

The EMMA can be applied for individual storm events to quantitatively evaluate the contribution of each solutions component. The source waters are called 'end members'. The tracer concentrations of end members are more extreme than stream water since streamwater is a mixture of these sources (Fig. 6). In order to apply EMMA, (1) tracers should be conservative, (2) sources should be significantly different in tracer concentrations, (3) unmeasured sources must have same concentration with known sources or don't contribute significantly, and (4) the sources should maintain a constant concentration. Typical source waters are those from organic rich soil horizon, hillslope groundwater, valley bottom groundwater, throughfall, and precipitation.

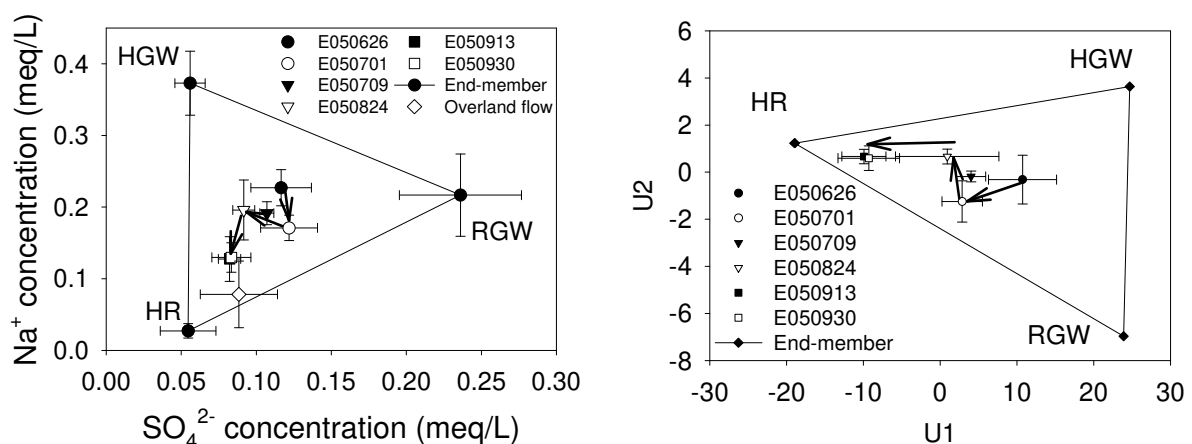


Fig. 6. Three-component mixing diagram for each storm event (left) and mixing diagram showing stream water evolution and end-member composition in U space during six storm events (right)

The hydrological characteristics of the six storm events observed during the summer of 2005 are summarized in Table 5. The maxima of precipitation intensity and discharge intensity were observed on 1 July 2005, which were 17.7 mm/10min. and 1.0 mm/10min., respectively. Stream discharge as a proportion of total precipitation ranged from 15 to 60% with an average of 30%. The maximum discharge rate also was observed in E050701, but associated with 5 days' antecedent precipitation.

The end-member mixing analysis (EMMA) with principal components analysis (PCA) was applied to each storm event to evaluate quantitatively the contribution of each water component (Christophersen and Hooper, 1992; Burns et al., 2001). Three-component mixing diagrams are shown in Fig. 6. Stormflow in E050626 lay near the groundwater end-member, and moved to soil water in E050701. Stormflow were also closer to that of groundwater through E050709 and E050824. After moving near groundwater, stormflow were closer to that of throughfall in E050913 and E050930. Stormflow solutes in E050913 and E050930 were not significantly different from overland flow.

In E050913 and E050930, the values of water-filled porosity in the surface layer (0–0.1 m) was about 5% higher than the maximum observed during the previous storm events. This

higher water-filled porosity (as compared to prior storm events) led to a low water infiltration rate and an increase in the contribution of surface discharge. Previous studies suggested that a maintained precipitation expands the saturation zone and increases macropore flows in the forested catchment (e.g., McDonnell, 1990). Such macropore flows deliver new water in which dissolved ion concentrations are low because of the short contact time with soil and bedrock (Burns et al., 1998). The calculated mean residence time of water based on the  $^{35}\text{S}$  analysis varied with changing water regime in the study area, ranging from 20 to 40 days during the summer monsoon period (Kim et al., 2009). Especially, for the stream water sample taken on 15 September when the surface runoff increased due to the storm event, the mean residence time of water also decreased abruptly (Kim et al., 2009; Fig. 5).

	E050626	E050701	E050709	E050824	E050913	E050930
Observed period	26–28, Jun.	1–3, Jul.	9–10, Jul.	24–26, Aug.	13–15, Sep.	30, Sep.–2, Oct.
Total precipitation (mm)	160.5	104.0	40.5	83.5	85.5	87.0
Max. precipitation intensity (mm/10min)	11.1	17.7	2.5	4.5	7.5	2.5
Total discharge (mm)	23.6	61.5	11.5	22.8	18.1	29.1
Max. discharge intensity (mm/10min)	0.32	1.05	0.06	0.14	0.28	0.16
Total discharge / Total precipitation (%)	15	60	28	27	21	33
Antecedent precipitation (5 days)	0.0	161.9	1.3	1.5	7.0	1.0
Antecedent precipitation (10 days)	1.3	161.9	154.3	19.5	7.0	43.5

Table 5. Hydrological characteristic of storm events in 2005

### 3. Dynamics of water and dissolved materials in forest soils

The dynamics of water in the soil layer are important for the understanding of water storage and dissolved material fluxes in a forest catchment. In the field measurement, an intensive monitoring is useful using a precise multiplex Time Domain Reflectometry system to capture and characterize variation patterns of soil moisture on a steep hillslope. Here, we introduce the methods for estimating the water and dissolved material flux in soils with tensiometer and water table fluctuations.

#### 3.1 Estimation of soil water and dissolved material flux using a tensiometer

Tensiometer consists of a pressure transducer which measures the pressure (when saturated) or tension (when unsaturated) that the soil moisture exerts on a column of water, a porous cup which is in contact with the soil water at the measurement level, and a water body with a PVC pipe. According to Kim (2003), the one-dimensional, vertical water flow equation for unsaturated soil in a compartment can be written as:

$$Q_{in} = Q_{out} - E + \Delta W \quad (2)$$

where  $Q_{in}$  and  $Q_{out}$  are input and output of water to and from the compartment, respectively,  $E$  is the evapotranspiration, and  $\Delta W$  is the change of water content in the compartment during the period. For example,  $Q_{in}$  in the 0-10 m soil compartment can be obtained from the throughfall measurement, and  $\Delta W$ ,  $E$  by direct observations. The calculated  $Q_{out}$ , in turn, becomes  $Q_{in}$  for the 0.1-0.2 m soil compartment. Therefore, the equation can be used to calculate the water flux through a series of compartments up to 1.0 m soil depth.

$E$  can be calculated from temporal variations of evapotranspiration (Suzuki, 1980).

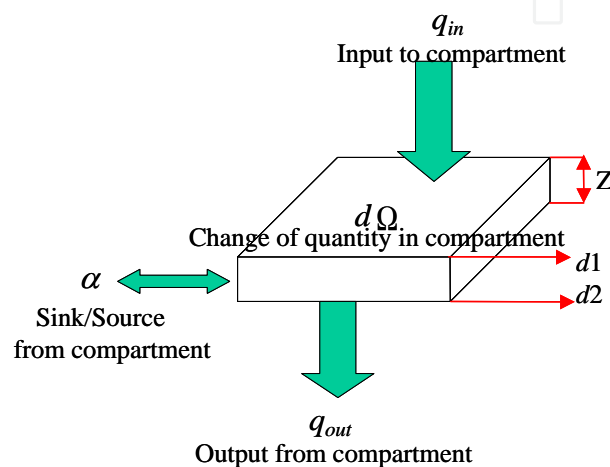


Fig. 7. Calculation of dissolved material flux in soil compartment (Kim, 2003)

$$E_{d1-d2} = cE \quad (3)$$

where  $E_{d1-d2}$  is evapotranspiration in soil depth from  $d1$  to  $d2$ ,  $E$  is the total evapotranspiration from the entire soil column, and  $c$  is the proportion of  $E_{d1-d2}$  to  $E$ . For example,  $c$  in the 0-0.1 m soil compartment (if the total soil depth is 1.0 m) during time  $t$  is calculated from the change of water content by using equation (4).

$$c = \frac{(\theta_{0-10}^{t+\Delta t} - \theta_{0-10}^t)}{(\theta_{0-10}^{t+\Delta t} - \theta_{0-10}^t) + (\theta_{10-20}^{t+\Delta t} - \theta_{10-20}^t) + (\theta_{20-30}^{t+\Delta t} - \theta_{20-30}^t) + (\theta_{30-50}^{t+\Delta t} - \theta_{30-50}^t) + (\theta_{50-100}^{t+\Delta t} - \theta_{50-100}^t)} \quad (4)$$

$\Delta W$  can be calculated from the change of water content, which is derived from the relationship between  $\theta$  and  $\psi$  (Kosugi, 1994; Kosugi, 1996).

$$\Delta W = \left( \theta_{(d1+d2)/2}^{t+\Delta t} - \theta_{(d1+d2)/2}^t \right) \cdot Z \quad (5)$$

where  $\theta_d^t$  is water content during time  $t$  at soil depth  $(d1+d2)/2$ , and  $Z$  is soil thickness.

Dissolved ions and compounds in soils move with water infiltration processes. Therefore, dissolved material flux is calculated by multiplying dissolved material concentration with the water flux. The calculation method of dissolved material flux is described in Fig. 7. The dissolved material flux is calculated from the change of quantity in a compartment. The

sink/source ( $a$ ) property of the compartment can be estimated from  $q_{in}$ ,  $q_{out}$  and the change of quantity in the compartment ( $d\Omega$ ), such as:

$$a = d\Omega - (q_{in} - q_{out}) \quad (6)$$

where  $d\Omega$  is calculated from the concentration of dissolved materials and water content.

$$d\Omega = \left( \theta_{(d1+d2)/2}^{t+\Delta t} \cdot S_{(d1+d2)/2}^{t+\Delta t} - \theta_{(d1+d2)/2}^t \cdot S_{(d1+d2)/2}^t \right) / Z \quad (7)$$

where  $S_{(d1+d2)/2}^t$  is the dissolved material concentration during time  $t$  at soil depth  $(d1+d2)/2$ . The equation (7) indicates the change of dissolved material budget in the soil compartment during time  $t$ . Moreover,  $q_{in}$  and  $q_{out}$  at depth  $d$  can be described as:

$$q_{in} = (f_{d1}^t + f_{d1}^{t+\Delta t}) / 2 \cdot \Delta t \quad (8)$$

$$q_{out} = (f_{d2}^t + f_{d2}^{t+\Delta t}) / 2 \cdot \Delta t \quad (9)$$

where  $f_{d1}^t$  is dissolved material flux at soil depth  $d1$  during time  $t$ .

### 3.2 Estimation of water infiltration rate using a water table fluctuation

The water infiltration rate can be calculated indirectly from the groundwater recharge rate. To estimate the water infiltration rate, the groundwater recharge rate from the water table fluctuation can be calculated as follows (Moon et al., 2004):

$$\alpha = \frac{\Delta h}{\sum P} \times S_y \quad (10)$$

where  $\alpha$  is the recharge rate,  $h$  is the change of groundwater level,  $P$  is precipitation, and  $S_y$  is the specific yield. On specific conditions, groundwater recharge rate may practically represent the infiltration rate. We can also estimate the dissolved material flux, such as dissolved organic carbon (DOC) by multiplying groundwater recharge rate with the measured concentration. This technique has been applied to the headwater region in the Gwangneung catchment, and its reliability has been critically evaluated by comparing with other methodologies. The uncertainty of this technique is largely due to the measurement error of specific yield ( $S_y$ ) caused by the heterogeneity of geologic materials, and other factors influencing the water table fluctuation such as changes in atmospheric pressures, air entrapment during the infiltration of water, irrigation, and pumping (Choi et al., 2007).

According to the results from the water infiltration rates, 0.44 t-C ha<sup>-1</sup> DOC was infiltrated into the soil from late June to early October in 2005, which represented approximately 8% of the stored carbon in the forest floor (5.6 t-C ha<sup>-1</sup>; Lim et al., 2003) and 30 to 50% of NEE (-0.84 to 1.56 t-C ha<sup>-1</sup> yr<sup>-1</sup>; Kwon et al., 2010) (Fig. 8). These results indicate that a considerable amount of decomposed organic matter is stored in the soil through water movement processes. If most of the infiltrated DOC were to accumulate as soil organic carbon in the shallow soil and to be decomposed in the deep soil, then 0.5% of the soil carbon (92.0 t-C ha<sup>-1</sup>; Lim et al., 2003) would be retained from DOC during the summer monsoon (Fig. 8).

While these values seem to be relatively small, soil organic carbon can be accumulated in the mineral soil for an extended period (e.g., Michalzik et al., 2003); potentially making the 0.5% of soil carbon retained from DOC during the summer monsoon an important component of the forest carbon budget to consider (e.g., Battin et al., 2009).

Based on these estimates of NPP ranging from 4.3 to 5.8 t C ha<sup>-1</sup> yr<sup>-1</sup>, the observed amount of total DOC and POC effluxes is roughly 2% of the annual NPP – a small but non-negligible amount in terms of net ecosystem carbon exchange (NEE). Considering the averaged NEE of -0.84 t C ha<sup>-1</sup> yr<sup>-1</sup> (negative sign indicates net uptake of carbon by the forest; Kwon et al., 2010), approximately 10% of NEE would escape from this forest catchment as DOC and POC (Fig. 8). Our results further indicate that 50 and 80% of the respective annual DOC and POC effluxes were transported out of this forest catchment during the summer monsoon period.

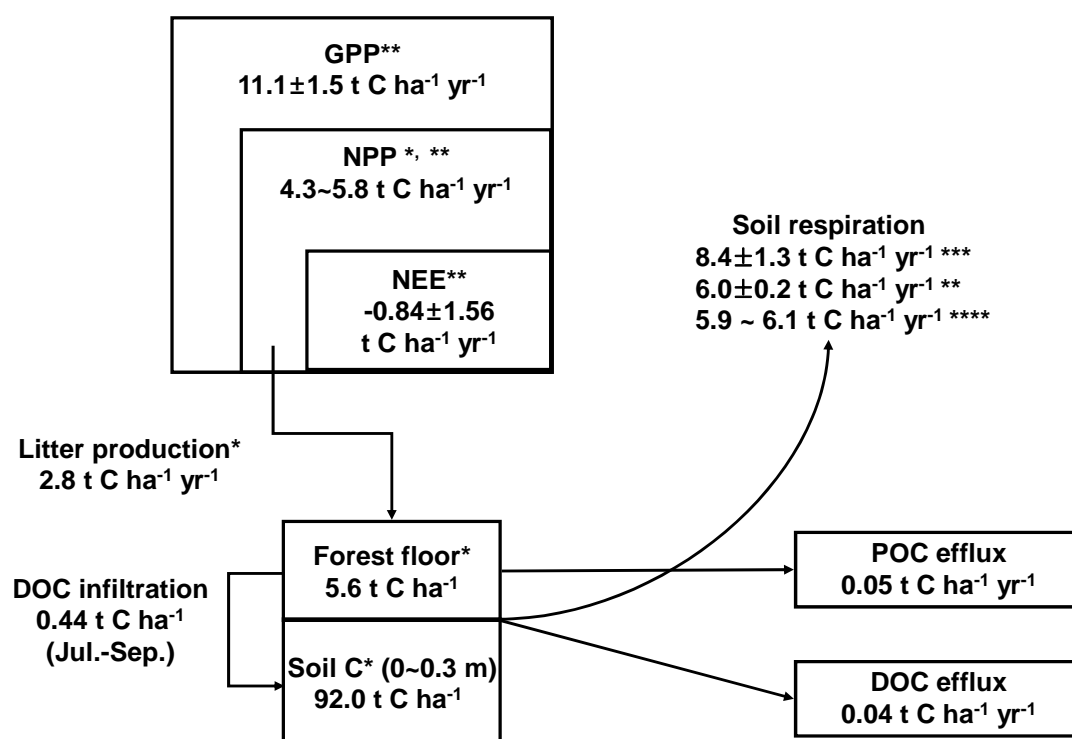


Fig. 8. The contribution of DOC and POC to the carbon budget in the Gwangneung deciduous forest catchment. \* Lim et al. (2003; observation periods 1998 to 1999), \*\* Kwon et al. (2010; observation periods 2006 to 2008), \*\*\* Chae (2008; observation periods 2001 to 2004). The difference of soil respiration is due to difference of observation periods and methods. Modified from Kim et al., 2010.

### 3.3 Adsorption of DOC in forest soil

Many field studies have shown that the concentration of DOC in soil water significantly decreases with increasing soil depth (Fig. 9). It is generally assumed that adsorption of DOC to the surface of mineral soil is important than decomposition in reducing DOC concentrations. Various sorption mechanisms have been reported, including anion exchange, cation bridging, physical adsorption, etc. (Jardine et al., 1989; Gu et al., 1994; Edwards et al., 1996; Kaiser and Zech, 1998a; Kaiser and Zech, 1998b). These DOC sorptions are irreversible under natural soil conditions (Gu et al., 1994). Because Fe and Al oxides are

the most important sources of variable charge in soils (Jardine et al., 1989; Moore et al., 1992; Kaiser and Zech, 1998a), DOC adsorption can be related quantitatively to the Fe and Al oxide contents of soils (Moore et al., 1992). The proportion of clay in mineral soil is also an important factor for DOC adsorption. DOC concentrations in catchment runoff are negatively correlated with the clay contents of soils in the catchment. The adsorption process is relatively rapid, which completed within 2 to 12 hours (Kaiser and Zech, 1998b). The effect of pH on the adsorption of DOC in forest soil is also important. Tipping and Woof (1990) calculated that an increase in soil pH by 0.5 units would lead to an increase by about 50% in the amount of mobilized organic matter. Nodvin et al. (1986) also calculated the reactive soil pool of DOC under various pH conditions.

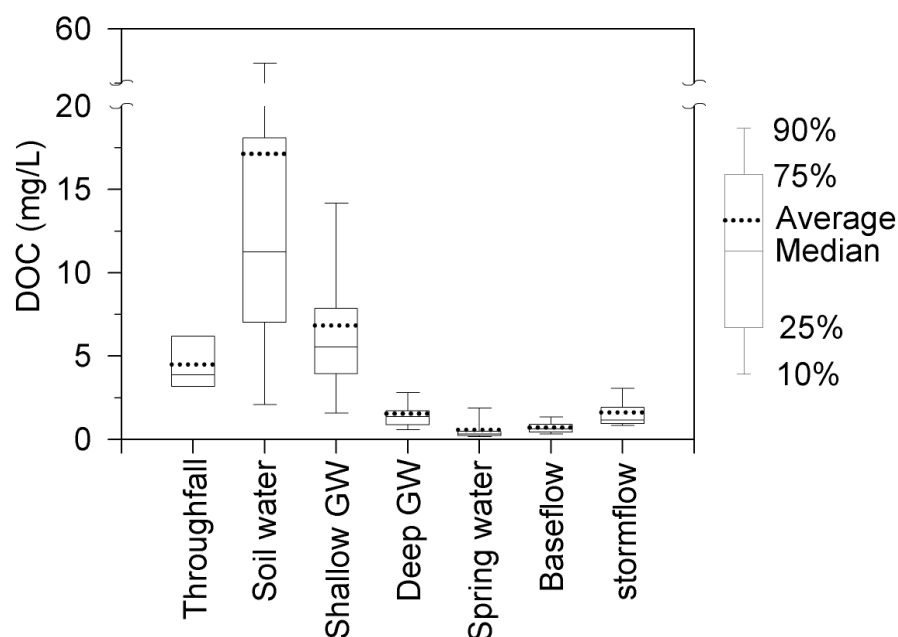


Fig. 9. Spatial variations in the concentrations of DOC of throughfall, soil water, shallow groundwater (0.5 m), deep groundwater (0.8-1.0 m), spring water, and baseflow, with respect to and stormflow (Kim et al., 2007b)

### 3.4 Temporal and seasonal change of DOC export from temperate forest catchment

Typical temporal variations in DOC concentrations during storm events are shown in Fig. 10. With the onset of heavy precipitation, DOC concentration in streamwater increases significantly, and after the precipitation ceased, DOC concentrations returned to pre-storm levels. The results from the hydrograph separation during storm events indicated that a large amount of water discharged through surface and subsurface soil layers (Fig. 6). DOC concentration in the surface soil is higher than the deep soil and the groundwater (Fig. 9). The Storm event leads to the increase in the surface runoff with a high DOC concentration. During the baseflow period, most stream waters flow out from the groundwater with a low DOC concentration (Fig. 11). These results indicate that hydrological processes strongly affect the DOC export and thereby the carbon budget in the catchment.



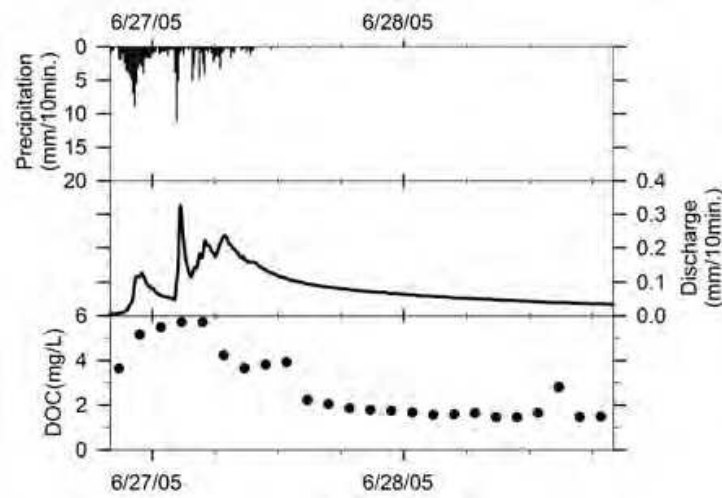


Fig. 10. Precipitation, stream discharge and temporal variations of DOC concentration in streamwater during storm event (Kim et al., 2007b)

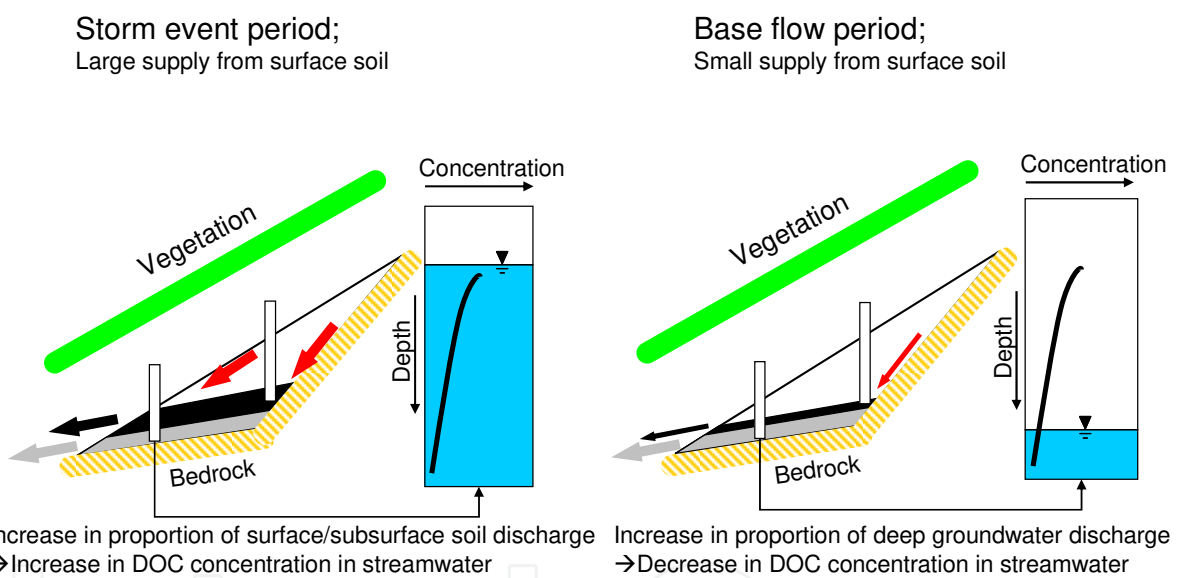


Fig. 11. Precipitation, stream discharge and temporal variations of DOC concentration in streamwater during storm event (Kim et al., 2007b)

Forest treatments		Throughfall (mm)	Stemflow (mm)	Interception loss (mm)	Interception loss (%)	Rainfall (mm)
<i>Pinus</i>	Not practice	85.8	6.5	74.0	44.6	166.0
<i>Koraiensis</i>	Practiced	100.2	7.5	58.3	35.1	
<i>Abie</i>	Not practice	90.1	7.8	68.1	41.0	
<i>holophylla</i>	Practiced	104.8	8.2	53.0	31.9	
	Deciduous forest	118.6	33.9	12.0	7.3	
	Mixed forest	115.3	17.6	33.1	19.9	

Table 6. Percentage of interception loss by forest practices and vegetation types

#### 4. Effects of forest managements on water cycle and quality

Change of the forest stand structure causes the modification of the hydrological characteristics because the components of water loss by interception and evapotranspiration change immediately, owing to the reduction of LAI. Moreover, physical and chemical properties of forest soil change in a long term.

Kim et al. (1993) conducted a research on the effects of site conditions in headwater stream on water storage of reservoirs on small-forested watersheds. The result shows that the water storage of the reservoirs during the dry season is positively correlated with the tree height, DBH, stand ages and crown closure, but negatively with understory coverage and drainage density.

The first change in hydrological components after forest practices is the interception loss from the tree canopy surface during rainfall. The amount of interception loss decreases after thinning and cutting. The rate of reduction for interception loss is correlated positively with the percentage of thinning and cutting of forest types. Table 6 represents the effects of forest practices and types on the percentage of interception loss. In the coniferous stand, the percentage of interception loss decreased to about 10% after forest practices. The mixed forest intercepted the rainfall in about half the amount of the coniferous, whereas the deciduous stand did in about one-sixths of that. Forest treatments increase not only interception loss but also discharge due to the reduction of loss components such as evapotranspiration. The amount of discharge after the forest practices during the dry season was increased by two and three-tenths of that, respectively, before the treatments.

Generally, forest soil has a filtering property like sponge and conserves soil and water resources. If forests, regardless of the types, are cut clearly, catastrophic amounts of soil and water are produced. Fig. 12 explains the effects of clear cutting on the peak flow in a small-forested catchment. The amount of peak flow in the clear-cut site increased to 78.3 mm compared to the controlled site during the rainfall of 400 mm.

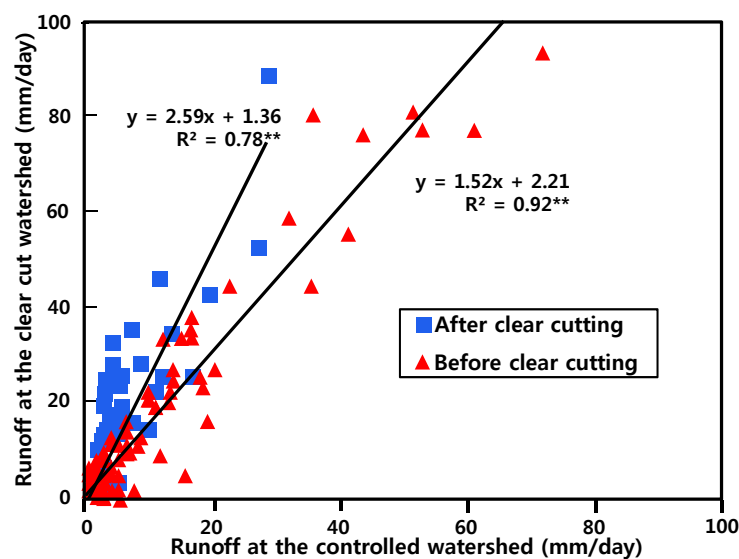


Fig. 12. Change of peak flow after clear cutting

Jeong et al. (1997) analyzed the influential factors on the electrical conductivity of stream and soil water in a small-forested watershed. They concluded that the electrical conductivity was correlated with the total amount of a cation and an anion in stream and soil water. Their results proposed that the amounts of  $\text{NO}_3^-$  and  $\text{Na}^+$  in the stream have a statistical significance for the electrical conductivity in streams and the amounts of  $\text{K}^+$  and  $\text{Ca}^{2+}$  and pH in soil water for the electrical conductivity in soil water.

Jeong et al. (1999a, b) further clarified the effect of forest management practices (thinning and pruning) on soil physical properties and water quality to obtain fundamental information on the facility of purifying water quality after forestry practices. They investigated the water quality of rainfall, throughfall, stemflow, and soil and stream water at the coniferous stands that consisted of *Abies holophylla* and *Pinus koraiensis*. The seasonal variation of water qualities of throughfall, stemflow and soil water were decreased after practices. Some researches supported the mesopore ratio on pore geometry of surface soil to be used as an index of the water retention capacity of forestlands.

Jeong et al. (2001a) investigated 23 parameters, including site conditions and soil properties to analyze the influencing factors of mesopore ratio on pore geometry of surface soil in coniferous stands. They found that the factors influencing the mesopore ratio (pF2.7) on the surface soil were macropore ratio (pF 1.6), slope, crown-cover rates, and thickness of F-layer, organic matter contents, and the growing stock. They concluded that crown-cover rates of stands should be controlled to be less than 80% for enhancing the water resource retention capacity in coniferous stands.

Jeong et al. (2001b) investigated fifteen factors, including site conditions and soil properties to analyze the influencing factors of mesopore ratio on a pore geometry of surface soil in deciduous stands. The factors influencing the mesopore ratio (pF2.7) on the surface soil were found to be the tree height, under vegetation coverage and organic matter contents of soil in deciduous stands. Hence, they concluded that the water resource retention capacity would be improved when under vegetation coverage was increased from 30 to 80%.

## 5. Upscaling of observation data through hydrological modeling

The rainfall-runoff process, which is an important component of the hydrological system, is very complex considering the large number of factors involved and their temporal and spatial distribution. Hydrological modelling is a suitable technique to represent the rainfall-runoff process in various symbolic or mathematical forms using known or assumed functions expressing the various components of a rainfall-runoff response (Ndiritu and Daniell, 1999). In the last half-century there have been hundreds of hydrological response models, each with their own attributes and shortcomings, developed by many different researchers. Furthermore, with the current rapid developments within computer technology and hydrology, the application of computer based hydrologic models is only likely to increase in the near future (Loague and Van der Kwaak, 2004).

The distributed hydrological models aim to better represent the spatio-temporal variability of hydrological characteristics governing the rainfall-runoff response at the catchment scale. One of the distributed hydrological models used commonly is TOPMODEL, which is a quasi-physically based semi-distributed hydrological model (Beven and Kirby, 1979; Beven et al., 1995; Beven, 1997; Beven, 2001; Beven and Freer, 2001a, b).

Most physically based distributed models have parameters which are effective at the scale of the computational elements. In order for a rainfall-runoff model to have practical utility or be useful for hypothesis testing, it is necessary to select appropriate values for the model parameters. Unfortunately, it is not normally possible to estimate the effective values of parameters by either prior estimation or measurement, even given intensive series of measurements of parameter values. Therefore, parameter values must be calibrated for individual applications (Refsgaard and Knudsen, 1996; Refsgaard, 1997; Freer, 1998; Beven, 2001).

In general, the process of parameter calibration has involved some form of determination of a parameter set that gives a simulation that adequately matches the observation. However, many calibration studies in the past have revealed that while one optimum parameter set could often be found, there would usually be a multitude of quite different parameter sets that can produce almost equally good simulation results. Recognition of multiple acceptance parameter sets results in the concept of equifinality of parameter sets (Beven and Freer, 2001b; Beven, 2002; Freer et al., 2003). In addition, in the general case for rainfall-runoff modelling with multiple storm sequences, it might be difficult to assess model performance using a single likelihood measure, because the form of the distribution of uncertain predictions varies markedly over the range of streamflow and the appropriate error structure might vary with both of type of data and the model parameter set (Freer et al., 2003). It may often be the case that the available data are not adequate to allow identification of complex models and/or that a single performance measure (objective function) is not adequate to properly take into account the simulation of all the characteristics of a system used. Thus, the multi-criteria or multi-objective methods using multiple objective functions or other data in addition to rainfall-runoff data may allow more robust analyses of models, and aid hypothesis testing of competing model structures (Gupta et al., 1999; Beven, 2001; Madsen et al., 2002; Freer et al., 2003).

The multi-criteria performance measures based on the concept of equifinality of behavioral model simulations were used for calibration of the rainfall-runoff model, TOPMODEL at natural deciduous forest in South Korea. Totally 100,000 parameter sets uniformly sampled by Monte Carlo Simulations from the ranges for each TOPMODEL parameters, and hourly stream flow and rainfall data observed from April to October, 2005 in the deciduous forest catchment located in the Gwangnung experimental forests were used for model calibration.

The performance of each parameter set was evaluated and identified with 6 different performance measures against behavioral acceptance thresholds defined for each performance measure, and the results were analyzed focused on the variability and relationship between the behavioral parameter distributions according to the definitions of performance measures.

The results demonstrate that there are many acceptable parameter sets scattered throughout the parameter space, all of which are consistent in some sense with the calibration data, and the range of model behavior for each parameter varied considerably between the different performance measures. Sensitivity was very high in some parameters, and varied depending on the kind of performance measure (Fig. 13). Compatibilities of behavioral parameter sets between different performance measures also varied, and a very small minority of parameter sets could produce reliable predictions regardless of the kind of performance measures (at least, for the performance measures used in this paper).

Especially, the results indicate that using a single performance measure for the calibration of a hydrological model may lead to an increase in model uncertainty. Therefore, careful consideration should be given to the choice of performance measure appropriate to the characteristics of used model and data and the purpose of study.

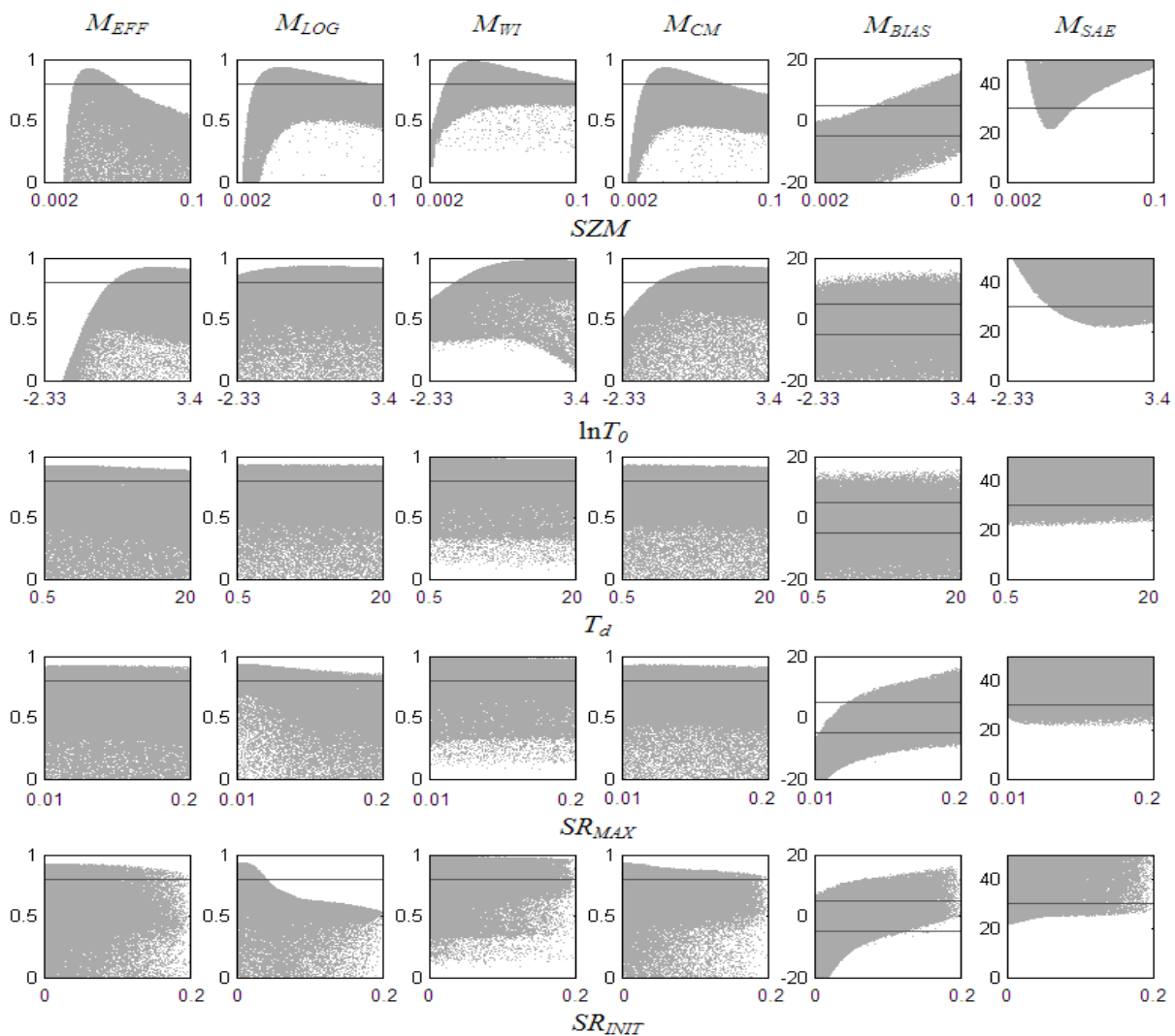


Fig. 13. Scatter plots of likelihood values for TOPMODEL parameters from Monte Carlo simulations of the deciduous forest catchment conditioned on the 2005 discharge period using six performance measures. Each dot represents one simulation with a likelihood weight calculated by a given performance measure, and horizontal lines mean thresholds identifying behavioural parameter sets for each performance measure; dots over the line (in cases of  $M_{EFF}$ ,  $M_{LOG}$ ,  $M_{WI}$  and  $M_{CM}$ ), dots between both lines (in case of  $M_{BIAS}$ ) and dots below the line (in case of  $M_{SAE}$ ) are classified as behavioural simulations.

Differences in the behavioral parameter distributions according to the performance measures may be directly caused by the definitions of performance measures. However, it also should be considered that the effects of model nonlinearity, covariation of parameter values and errors in model structure, input data or observed variables may be taken into account in the nonlinearity of the response of acceptable model.

The performance of the parameter set can be used to produce the likelihood-weighted marginal parameter distributions for individual parameters, and the likelihood weighted model simulations can be used to estimate prediction quantiles in a way that allows that different models may contribute to the ensemble prediction interval at different time steps and that the distributional form of the predictions may change from time to time step (Fig. 14).

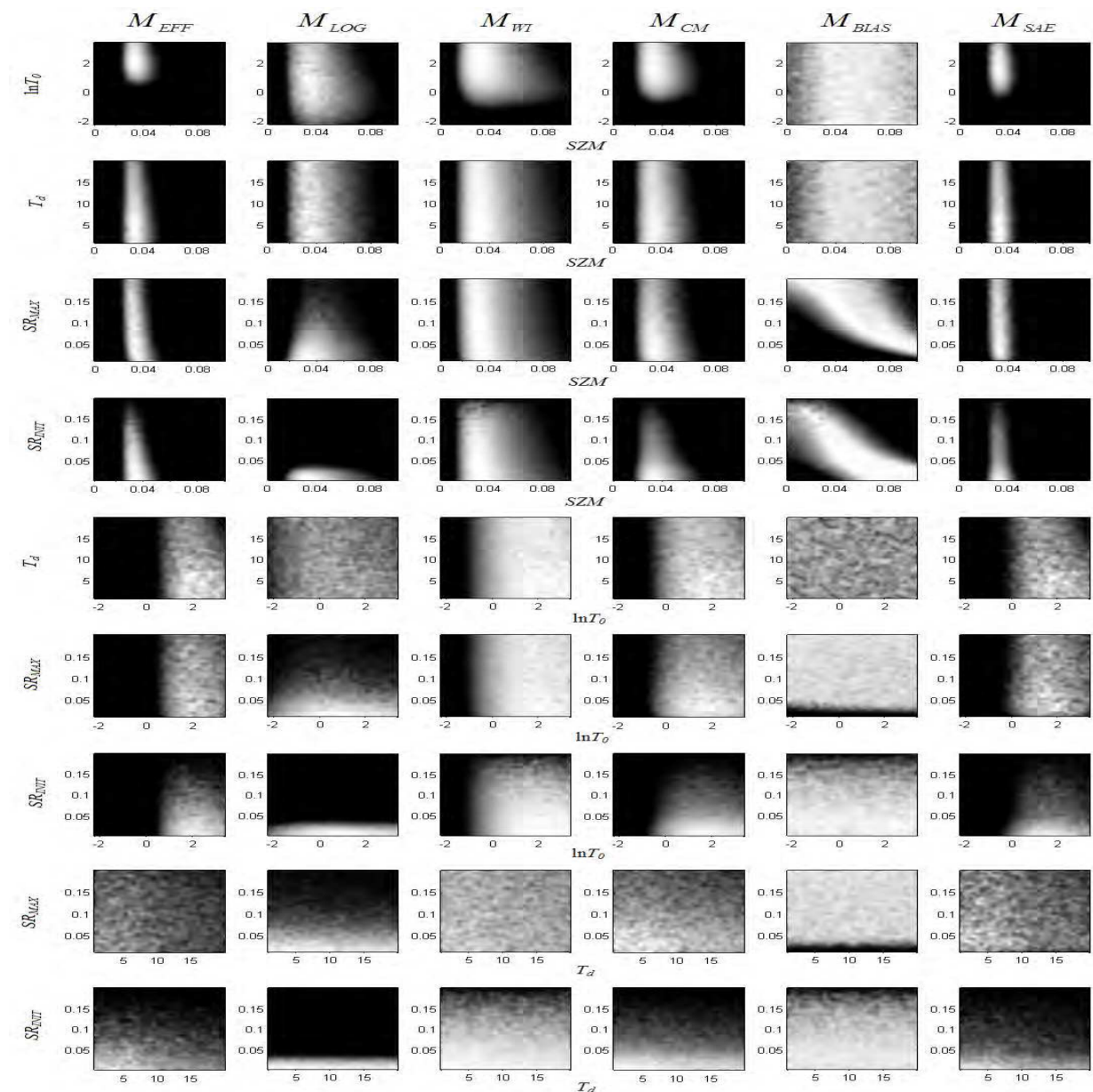


Fig. 14. Likelihood response surfaces between the major parameters of TOPMODEL, conditioned on the 2005 discharge period of the deciduous forest catchment (Behavioural parameter sets with higher model performance are in the white zone.)

## 6. Conclusions

The ecohydrological and biogeochemical studies have proposed a major scientific question: What is the role of hydrology in the carbon budget of complex forest catchment and how will

it change in the hydrologic cycle in monsoon Asia and influence the forest carbon budget? (Kim et al., 2006) To properly answer this question, some of the most fundamental aspects in catchment hydrology need to be clarified i.e., (1) How much water is stored in the catchments? (2) What flowpaths does water take to the stream? (3) How long does water reside in catchments? (4) How can we scale or transfer our observations to other catchments? Despite decades of dedicated scientific efforts on these fundamental questions, it is still difficult to find a robust interpretation even for some basic hydrological processes such as discharge and runoff. The up to date results showed that the geophysical and meteorological conditions greatly affect the hydrological processes (Hooper et al., 1990; Elsenber et al., 1995; Katsuyama et al., 2001; McGlynn and McDonnell, 2003; Kim et al., 2010).

To understand carbon cycling in this catchment better, it is necessary to estimate the annual accumulation and movement of water and DOC in the soil. The organic carbon has been continuously discharged from terrestrial ecosystems of river basin. This organic carbon will contribute for an important sink for carbon through burial in coastal sea sediments or floor. These missing values have to consider for estimation of carbon budget in terrestrial ecosystems. Our results suggest that storm events during summer monsoon (including the typhoon season) are important to estimate flow paths of water and carbon budget in a Korean forested catchment and East Asia. The seasonally concentrated precipitation increases the surface runoff, when the infiltration capacity of the soil decreases during summer monsoon. The outbreak of surface runoff reduced the mean residence time of water in the catchment, and increased DOC export from the surface soil layer. The precipitation also plays an important role in infiltration processes of dissolved material. The precipitation patterns and hydrological processes strongly affect the carbon cycling in the Korean temperate forest during summer monsoon. The increasing occasions of heavy precipitation may not lead to the simultaneous increase of available water resources in the catchment due to the shortening of the water residence time. However, the heavy precipitation will clearly increase material discharge such as DOC. Therefore, the effect of monsoon climate on water and carbon cycling in forest catchment should be critically evaluated on the basis of improved understanding of catchment hydrological and biogeochemical processes.

Our understandings in water and carbon cycling obtained from the hydro-biogeochemical approaches are limited due to the prescribed spatial scale of the measurements. The scaling issues are implicitly built into our field measurements and model representations (Kim et al., 2006). The information provided in this chapter should be carefully considered in modelling formulations at the hydrologic catchment and grid scales of ecohydrological/biogeochemical models and satellite image analyses. Such efforts should provide insights as to how various information is transferred across scales, and hence on how to simplify and aggregate measurements, models and satellite products. Future research must be focused on how to make measurements at scales that are appropriate for parameterization and model validation, and how to make the scales of modeling and satellite algorithm converge with those of field measurements (Kim et al., 2006).

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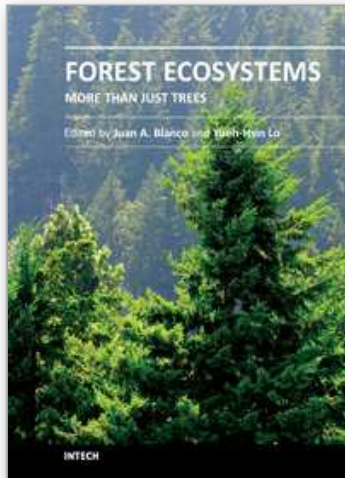
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## **Forest Ecosystems - More than Just Trees**

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The common idea for many people is that forests are just a collection of trees. However, they are much more than that. They are a complex, functional system of interacting and often interdependent biological, physical, and chemical components, the biological part of which has evolved to perpetuate itself. This complexity produces combinations of climate, soils, trees and plant species unique to each site, resulting in hundreds of different forest types around the world. Logically, trees are an important component for the research in forest ecosystems, but the wide variety of other life forms and abiotic components in most forests means that other elements, such as wildlife or soil nutrients, should also be the focal point in ecological studies and management plans to be carried out in forest ecosystems. In this book, the readers can find the latest research related to forest ecosystems but with a different twist. The research described here is not just on trees and is focused on the other components, structures and functions that are usually overshadowed by the focus on trees, but are equally important to maintain the diversity, function and services provided by forests. The first section of this book explores the structure and biodiversity of forest ecosystems, whereas the second section reviews the research done on ecosystem structure and functioning. The third and last section explores the issues related to forest management as an ecosystem-level activity, all of them from the perspective of the other parts of a forest.

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