

1 Title : Hydro-Biogeochemical Approaches to Understanding of Water and Carbon
2 Cycling in the Gwangneung Forest Catchment

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20 한국농림기상학회지 투고규정에 의거 본 논문을 심사용으로 제출합니다.

1 **Hydro-Biogeochemical Approaches to Understanding of**
2 **Water and Carbon Cycling in the Gwangneung Forest**
3 **Catchment**

4
5 **Abstract**

6
7 The information on flowpath, storage, residence time, and interactions of water
8 and carbon transport in a catchment is the prerequisite to the understanding and
9 predicting of water and carbon cycling in the mountainous landscapes of Korea. In this
10 paper, along with some up-to-date results, we present the principal methods that are
11 currently used in HydroKorea and CarboKorea research to obtain such information.
12 Various catchment hydrological processes have been examined on the basis of the water
13 table fluctuations, the end-member mixing model, the cross correlation analysis, and
14 cosmogenic radioactive isotope activity. In the Gwangneung catchment, the
15 contribution of surface discharge was relatively large, and the changes in the amount,
16 intensity and patterns of precipitation affected both the flowpath and the mean residence
17 time of water. Particularly during the summer monsoon, changes in precipitation
18 patterns and hydrological processes in the catchment influenced the carbon cycle such
19 that the persistent precipitation increased the discharge of dissolved organic carbon
20 (DOC) concentrated in the surface soil layer. The improved understanding of the
21 hydrological processes presented in this report will enable a more realistic assessment
22 of the effects of climate changes on the water resource management and on the carbon
23 cycling in forest catchments.

1

2 *Key words* : Gwangneung forest catchment, Hydro-biogeochemistry, Monsoon, Carbon

3 and water cycling, DOC

4

1 I . Introduction

2 The water and carbon cycles in forest catchments are important elements for
3 understanding the impact of global environmental changes on terrestrial ecosystems.
4 Various theories have been suggested to better understand water discharge (Horton,
5 1933; Betson, 1964; Kirkby, 1978; Anderson and Burt, 1991; Kim *et al.*, 2003) and its
6 effect on carbon efflux processes from forest catchments (McGlynn and McDonnell,
7 2003; Kawasaki *et al.*, 2005; Schulze, 2006). Most of the results indicated that the
8 hydrological flowpaths are important in carbon dynamics within the forest catchments.
9 In Korea, more than 50% of the annual precipitation falls in the summer monsoon
10 season, which quickly discharges to the ocean due to the steep slopes and short river
11 lengths (< 500 km). Therefore, the water regime in the catchment undergoes drastic
12 changes with recurring wet and dry seasons, which makes it difficult to interpret and
13 predict hydrological processes and subsequently their effect on carbon cycling.

14 The transport of terrestrial carbon into streams, rivers and eventually the oceans is
15 an important link between terrestrial and oceanic carbon cycles (Ludwig *et al.*, 1996;
16 Warnken and Santschi, 2004). As compared to the terrestrial carbon sinks (1.9 Gt-C/yr;
17 Prentice *et al.*, 2001), the organic carbon transport from terrestrial ecosystems to oceans
18 is 0.4 Gt-C/yr (Table 1), representing a subordinate but substantial component of the
19 ecosystem carbon balance.

20
21 [Table 1]
22

23 The ecohydrology group of HydroKorea and CarboKorea projects has proposed a
24 major scientific question: What is the role of hydrology in the carbon budget of

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

12 There have been few studies that involve intensive ecohydrological measurements
13 for the comprehensive understanding of hydrological processes and their effect on
14 carbon cycling in Korean forest catchment. Therefore, we have implemented a
15 comprehensive ecohydrological measurement system at the Gwangneung Supersite in
16 Korea. Most importantly, high quality long-term data of hydrological and
17 meteorological conditions have been collected, which may be also important in
18 monitoring global environmental changes and their effects. The study was also designed
19 based on a nested watershed concept (smaller catchments are nested in successively
20 larger catchments) to investigate how catchment processes change as scale varies. A
21 more detailed description of the study site and the experiment design in the
22 Gwangneung Supersite can be found in Lee *et al.* (this issue). In this paper, we
23 introduce the concepts and techniques that were implemented to investigate the
24 movement of water and carbon in a forest catchment. We also briefly discuss

- 1 preliminary results and their implications for the interactions between hydrological and
- 2 biogeochemical processes in a catchment.
- 3

1 **II . Dynamics of water in forest soils**

2 The dynamics of water in the soil layer are important for the understanding of
3 water storage and dissolved material fluxes in a forest catchment. In the Gwangneung
4 catchment, an intensive monitoring is being conducted using a precise multiplex Time
5 Domain Reflectometry system to capture and characterize variation patterns of soil
6 moisture on a steep hillslope (Kim *et al.*, 2007). Here, we introduce the methods for
7 estimating the water and dissolved material flux in soils with tensiometer and water
8 table fluctuations.

10 **2.1. Estimation of soil water and dissolved material flux using a tensiometer**

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

$$16 \qquad Q_{in} = Q_{out} - E + \Delta W \qquad (1)$$

17 where Q_{in} and Q_{out} are input and output of water to and from the compartment
18 respectively, E is the evapotranspiration, ΔW is the change of water content in the
19 compartment during the period. For example, Q_{in} in the 0–10 cm soil compartment can
20 be obtained from the throughfall measurement, and ΔW , E by direct observations. The
21 calculated Q_{out} , in turn, becomes Q_{in} for the 0.1–0.2 m soil compartment. Therefore, the
22 equation can be used to calculate the water flux through a series of compartments up to
23 1.0 m soil depth.

1 E can be calculated from temporal variations of evapotranspiration (Suzuki
2 1980).

$$3 \quad E_{d1-d2} = cE \quad (2)$$

4 where E_{d1-d2} is the evapotranspiration at soil depth from $d1$ to $d2$, E is the total
5 evapotranspiration from the entire soil column, and c is the proportion of E_{d1-d2} to E . For
6 example, c in the 0–0.1 m soil compartment (if the total soil depth is 1.0 m) during time
7 t is calculated from the change of water content by using equation (3).

$$8 \quad c = \frac{(\theta_{0-10}^{t+\Delta t} - \theta_{0-10}^t)}{(\theta_{0-10}^{t+\Delta t} - \theta_{0-10}^t) + (\theta_{10-20}^{t+\Delta t} - \theta_{10-20}^t) + (\theta_{20-30}^{t+\Delta t} - \theta_{20-30}^t) + (\theta_{30-50}^{t+\Delta t} - \theta_{30-50}^t) + (\theta_{50-100}^{t+\Delta t} - \theta_{50-100}^t)}$$

9 (3)

10 ΔW can be calculated from the change of water content, which is derived from the
11 relationship between θ and ψ (Kosugi, 1994; Kosugi, 1996).

$$12 \quad \Delta W = (\theta_{(d1+d2)/2}^{t+\Delta t} - \theta_{(d1+d2)/2}^t) \cdot Z \quad (4)$$

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

16 [Fig. 1]

18 Dissolved ions and compounds in soils move with water infiltration processes.
19 Therefore, dissolved material flux is calculated by multiplying dissolved material
20 concentration with the water flux. The calculation method of dissolved material flux is
21 described in Fig. 1. The dissolved material flux is calculated from the change of
22 quantity in a compartment. The sink/source (α) property of the compartment can be
23 estimated from q_{in} , q_{out} and the change of quantity in the compartment ($d\Omega$), such as;

$$\alpha = d\Omega - (q_{in} - q_{out}) \quad (5)$$

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

$$d\Omega = \left(\theta_{(d1+d2)/2}^{t+\Delta t} \cdot S_{(d1+d2)/2}^{t+\Delta t} - \theta_{(d1+d2)/2}^t \cdot S_{(d1+d2)/2}^t \right) / Z \quad (6)$$

where $S_{(d1+d2)/2}^t$ is the dissolved material concentration during time t at soil depth $(d1+d2)/2$. The equation (6) 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

$$q_{in} = (f_{d1}^t + f_{d1}^{t+\Delta t}) / 2 \cdot \Delta t \quad (7)$$

$$q_{out} = (f_{d2}^t + f_{d2}^{t+\Delta t}) / 2 \cdot \Delta t \quad (8)$$

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

2.2. 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):

$$\alpha = \frac{\sum h}{\sum P} \times S_y \quad (9)$$

where α 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

1 error of specific yield (S_y) caused by the heterogeneity of geologic materials, and other
2 factors influencing the water table fluctuation such as changes in atmospheric pressures,
3 air entrapment during the infiltration of water, irrigation, and pumping (Choi *et al.*, this
4 issue).
5

1 **III. Hydrological processes and the mean residence time of water in a** 2 **forest catchment**

3 Understanding the hydrological processes that control the movement of water in a
4 forest catchment, in most cases, cannot be simply derived from physical characteristics.
5 The physical information in hydrometric measurements provides insight on boundary
6 conditions for hydrological systems, but they have limited utility in determining the
7 residence time of water or specific flow pathways. In this section, we introduce the
8 methods to estimate hydrological flowpaths during storm events and the mean residence
9 time of water using hydro-biogeochemical approaches.

10 11 **3.1. Understanding of flowpaths using a hydrograph separation**

12 Runoff may be separated into three components: surface runoff, interflow and
13 groundwater runoff. The solute concentrations in each component may differ
14 significantly depending on the runoff sources. Therefore, an accurate observation of the
15 temporal changes in chemical composition of streamwater is fundamental to understand
16 the runoff generating processes.

17 The characterization of flowpaths in forested catchments has been elusive because
18 of the difficulties in measuring subsurface flow. Forested catchments are spatially
19 complex and the subsurface flow is invisible. Hence, one can only infer the movement
20 and mixing of water with the help of natural tracer elements that they carry (Pinder and
21 Jones, 1969). Using various tracers, end-member mixing analysis has been widely
22 conducted to elucidate pathways and relevant hydrological processes in a number of
23 catchments (e.g. Hooper *et al.*, 1990; Christophersen *et al.*, 1990; Elsenbeer *et al.*, 1995;
24 Katsuyama *et al.*, 2001).

1

2

[Fig. 2]

3

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

14

15

[Fig. 3]

16

17 Here we introduce the application of EMMA in the Gwangneung catchment
18 during storm events. The concentrations of SO_4^{2-} and Na^+ were selected for the analysis
19 because this combination provides the best separation of sources (Katsuyama *et al.*,
20 2001). For the period from April to October 2005, Fig. 3 shows the mean and variations
21 in the concentrations of SO_4^{2-} and Na^+ of throughfall, soil water, shallow groundwater
22 (0.5 m), deep groundwater (0.8–1.0 m), spring water, baseflow and stormflow measured
23 in this study. Average concentrations of SO_4^{2-} and Na^+ were highest in soil water and
24 deep groundwater, respectively. The shallow and the deep groundwater were not
25 different significantly in the concentrations of these solutes (Fig. 3). Therefore, the deep

1 groundwater could be treated as the first end-member of streamwater because Na⁺
 2 concentrations were highest possibly due to the accumulated effect of geochemical
 3 weathering. The throughfall can be treated as the second end-member because it is the
 4 juvenile component without being affected by catchment biogeochemical processes and
 5 is mixed directly into the streamwater. The precipitation reached on the forest floor
 6 infiltrates and flows downstream. Hence, we can select the soil water as the third
 7 end-member.

8 The contribution of each end-member was calculated by solving the following
 9 mass balance equations simultaneously (Hooper *et al.*, 1990; Katsuyama *et al.*, 2001) :

$$10 \quad f_a + f_b + f_c = 1 \quad (10)$$

$$11 \quad [\text{SO}_4^{2-}]_a f_a + [\text{SO}_4^{2-}]_b f_b + [\text{SO}_4^{2-}]_c f_c = [\text{SO}_4^{2-}]_{st} \quad (11)$$

$$12 \quad [\text{Na}^+]_a f_a + [\text{Na}^+]_b f_b + [\text{Na}^+]_c f_c = [\text{Na}^+]_{st} \quad (12)$$

13 where subscripts *a*, *b*, and *c* refer to the three different flowpaths; the subscript *st* refers
 14 to stream samples; *f* refers to the fraction of stream discharge contributed by each
 15 flowpath; and [SO₄²⁻] and [Na⁺] refer to the concentrations of each ion.

16 Fig. 4 shows precipitation, water-filled porosity, groundwater level, and stream
 17 discharge for a storm event. As expected, the water-filled porosity in the shallow soil
 18 layer increased more quickly and to a greater extent than in the deeper soil layer. The
 19 groundwater level at R1-G4 immediately responded to the precipitation. In E050626,
 20 the groundwater level in the beginning was 0.3 m below the surface, which gradually
 21 increased to the subsurface and then decreased.

22

23 [Fig. 4]

24

3.2. Time lags between precipitation and other hydrological components during storm events

We can quantify the time lag between the precipitation and the stream discharge or the soil moisture by calculating the cross correlation (C_{AB}), which measures the persistence of two signals (A and B) during the measurement period and is defined as (Stull, 1988):

$$C_{AB}(L) = \frac{\sum_{k=0}^{N-j-1} [(A_k - \overline{A_k})(B_{k+j} - \overline{B_{k+j}})]}{\left[\sum_{k=0}^{N-j-1} (A_k - \overline{A_k})^2 \right]^{1/2} \left[\sum_{k=0}^{N-j-1} (B_{k+j} - \overline{B_{k+j}})^2 \right]^{1/2}} \quad (13)$$

where L is the time lag ($=j\Delta t$), Δt is the measurement interval, $\overline{A_k} = \frac{1}{N-j} \sum_{k=0}^{N-j-1} A_k$ and

$$\overline{B_{k+j}} = \frac{1}{N-j} \sum_{k=0}^{N-j-1} B_{k+j}.$$

Fig. 5 shows the time lags simulated from the cross correlation analysis among stream discharge, precipitation, and soil moisture. If the time lag between precipitation and soil moisture was longer than the time lag between precipitation and stream discharge, we can presume that the surface runoff occurred by the high precipitation intensity. This phenomenon occurs when the rate of precipitation on a surface exceeds the rate at which water infiltrates the ground, and any depression storage has already been filled. This is called the Hortonian overland flow (Horton, 1933). This occurs more commonly in arid and semi-arid regions where rainfall intensities are high and the soil infiltration capacity is reduced because of surface sealing, or in paved areas. When the soil is saturated and the depression storage is filled, the precipitation will immediately generate surface runoff, which is called saturated overland flow. In the Gwangneung

1 catchment, the saturated overland flow was commonly observed especially during
2 summer monsoon.

3
4 [Fig. 5]
5

6 **3.3. Residence time of water in a forest catchment**

7 Various radioactive tracers have provided valuable information regarding
8 hydrological processes, such as mean residence time of water, flowpaths during storm
9 events, groundwater movement, and biogeochemical reactions occurring along the
10 flowpaths (Michel and Naftz, 1995; Shanley *et al.*, 1998; Sueker *et al.*, 1999). For
11 example, ^3H and ^{14}C have been widely used for determination of time scale of
12 hydrological processes (Matsutani *et al.*, 1993). However, these tracers are inadequate
13 for studying hydrological processes in small and headwater catchments with expected
14 time scales of a year or less because of their long half lives (decades to thousands of
15 years). In this study, we will introduce a short-lived cosmogenic radioactive isotope of
16 ^{35}S (half life = 87 days) for measuring the mean residence time of water in the
17 Gwangneung catchment.

18 The measured activity of ^{35}S in water can be expressed as an equation;

$$19 \quad C = C_0 e^{-\lambda t} \quad (14)$$

20 where C_0 is the initial ^{35}S activity, λ is the decay constant (0.0079655), t is the number
21 of days from the start of decay, and C is the measured ^{35}S activity.

22
23 [Fig. 6]
24

1 The ³⁵S activity in water provided information of the residence time of
2 atmospherically deposited sulfate. Biogeochemical reactions such as
3 adsorption/desorption in soil and groundwater are also important in affecting the
4 calculated residence time of water in a forested catchment. Assuming a conservative
5 response of sulfate in streamwater, the mean residence time of water was < 40 days
6 during the summer monsoon period in the Gwangneung catchment. However, the mean
7 residence time of water increased to around 100 days in the dry season with increasing
8 contribution of the base flow to the stream water (Fig. 6). These results demonstrate that
9 ³⁵S is useful in estimating the age of water exiting a small catchment where the time
10 scales of hydrologic processes are on the order of 1 year or less.

11

1 **IV. Effects of hydro-biogeochemical processes on soil carbon dynamics** 2 **in a forest catchment**

3 Biotic/abiotic factors affect carbon production/consumption in forest soils (Kalbitz
4 *et al.*, 2000). One of the most important abiotic factors in soil carbon dynamics is the
5 hydrological process. In this section, we review biogeochemical characteristics of forest
6 soil carbon, and discuss the effects of hydrological processes on soil carbon dynamics
7 based on the field data obtained from the Gwangneung forest catchment.

8 9 **4.1. Forms of carbon in soil-, ground-and stream water, and biogeochemical** 10 **reactions**

11 The main forms of carbon in stream water are particulated organic carbon (POC),
12 dissolved organic carbon (DOC), and dissolved inorganic carbon (DIC). The distinction
13 between DOC and POC is generally made on the basis of whether or not it passes
14 through filters with 0.45-0.50 μm pore size (Herbert and Bertsch, 1995). Soil water
15 contains varying amounts of DOC, which is originated from litter and other biomass.
16 Since most of the DOC in soil water consists of complex and high molecular weight
17 compounds, a general chemical definition of DOC is difficult to derive. Litterfall
18 represents the most important source of DOC and POC inputs to the forest catchment.
19 The turnover of DOC in soils is important as a major pathway of element cycling. DOC
20 is also a major controlling factor in soil formation (Dawson *et al.*, 1978; Kawasaki *et al.*,
21 2005). POC concentration in stream water is more affected by the surface runoff in a
22 forest catchment. The relationship between DOC concentration in stream water and
23 hydrological processes will be discussed in following sections.

24

4.2. Importance of DOC in soil carbon cycling

DOC in soils plays important roles in the biogeochemical cycling of carbon, nitrogen, phosphorus, and other nutrients in aquatic ecosystems, and in the transport of pollutants in soils (Kalbitz *et al.*, 2000; Perakis and Hedin, 2002).

[Fig. 7]

The conceptual model for soil carbon cycling is described in Fig. 7. The soil carbon is divided into heavy and light fraction (Herbert and Bertsch, 1995). Most of the soil carbon is the heavy fraction, which includes microbes and refractory organic compounds (Guggenberger *et al.*, 1994; Kalbitz *et al.*, 2000). The rest of the soil carbon is the light fraction, which is mainly originated from the litterfall and consists of labile organic compounds (Guggenberger *et al.*, 1994; Kalbitz *et al.*, 2000). Therefore, light fraction is dominant in organic or humic layers. The proportion of dissolved form in soil carbon is relatively small (< 0.1 %) compared to the heavy and light fractions. However, heterotrophic microbes mostly metabolize dissolved forms as their energy source (Sato and Seto, 1999; Kalbitz *et al.*, 2000; Wagai and Sollins, 2002). Dissolved form of carbon is used for cell formation, for microbial respiration, for desorption/adsorption, and for sedimentation from soil (Kalbitz *et al.*, 2000). The dissolved form is preserved in small quantities as a result of the balance among the dissolution of soil organic matter, the consumption by microbes, and the absorption/desorption to/from minerals. Therefore, the mass balance of DOC can be an important indicator for soil biogeochemical processes.

4.3. Adsorption of DOC in forest soil

1 Many field studies have shown that the concentration of DOC in soil water
2 significantly decreases with increasing soil depth (Fig. 8). It is generally assumed that
3 adsorption of DOC to the surface of mineral soil is important than decomposition in
4 reducing DOC concentrations. Various sorption mechanisms have been reported,
5 including anion exchange, cation bridging, physical adsorption, etc. (Jardine *et al.* 1989;
6 Gu *et al.*, 1994; Edwards *et al.*, 1996; Kaiser and Zech, 1998a; Kaiser and Zech, 1998b).
7 These DOC sorptions are irreversible under natural soil conditions (Gu *et al.*, 1994).
8 Because Fe and Al oxides are the most important sources of variable charge in soils
9 (Jardine *et al.* 1989; Moore *et al.*, 1992; Kaiser and Zech, 1998a), DOC adsorption can
10 be related quantitatively to the Fe and Al oxide contents of soils (Moore *et al.*, 1992).
11 The proportion of clay in mineral soil is also an important factor for DOC adsorption.
12 DOC concentrations in catchment runoff are negatively correlated with the clay
13 contents of soils in the catchment. The adsorption process is relatively rapid, which
14 completed within 2 to 12 hours (Kaiser and Zech, 1998b). The effect of pH on the
15 adsorption of DOC in forest soil is also important. Tipping and Woof (1990) calculated
16 that an increase in soil pH by 0.5 units would lead to an increase by about 50 % in the
17 amount of mobilized organic matter. Nodvin *et al.* (1986) also calculated the reactive
18 soil pool of DOC under various pH conditions.

19

20 [Fig. 8]

21

22 **4.4. DOC flux and accumulation in forest soils**

23 Kawasaki *et al.* (2005) calculated the DOC flux and accumulation in forest soil by
24 the multiplication of soil water flux and DOC concentration. They reported that DOC

1 removal rate is 87 % of the total incoming DOC flux to the surface mineral soil. As
2 described above, this result correlates with DOC adsorption mechanisms.

3 Lim *et al.* (2003) estimated that the annual net primary productivity (NPP) in
4 Gwangneung deciduous forest is about 4.3 t-C ha⁻¹ yr⁻¹. The annual DOC discharge
5 from Gwangneung deciduous forest catchment was estimated at about 0.04 t-C ha⁻¹ yr⁻¹
6 (Kim and Kim, 2006). Such a small magnitude of DOC discharge necessitates
7 estimations of retention and release of DOC within the soil. Based on the results from
8 water infiltration rate, an average of 0.45 t-C ha⁻¹ of DOC was infiltrated into the soil
9 from late June to early October. These results indicate that ~8 % of the litterfall carbon
10 (5.6 t-C ha⁻¹; Lim *et al.*, 2003) was stored in soil as DOC with water movement. If most
11 of the infiltrated DOC was preserved and accumulated as soil organic carbon in the
12 shallow soil, the 0.45 t-C ha⁻¹ of DOC that corresponds to ~0.5 % of the total soil
13 carbon (92.0 t-C ha⁻¹; Lim *et al.*, 2003) and ~10 % of the annual NPP would be retained
14 during summer monsoon. Although these values are small, it should be noted that DOC
15 is the most reactive component in the bulk soil organic carbon and therefore, its amount
16 may change drastically depending on the climatic and hydrological conditions.
17 Therefore, this 0.5 % of soil carbon retained as DOC can be considered as an important
18 component of the carbon budget in forest ecosystems.

20 **4.5. Temporal change of DOC concentration in streamwater during storm events**

21 Typical temporal variations in DOC concentrations during storm events are shown
22 in Fig. 9. With the onset of heavy precipitation, DOC concentration in streamwater
23 increases significantly, and after the precipitation ceased, DOC concentrations returned
24 to pre-storm levels. The results from the hydrograph separation during storm events
25 indicated that a large amount of water discharged through surface and subsurface soil

1 layers (Fig. 4). DOC concentration in the surface soil is higher than the deep soil and
2 the groundwater (Fig. 8). The Storm event leads to the increase in the surface runoff
3 with a high DOC concentration. During the baseflow period, most stream waters flow
4 out from the groundwater with a low DOC concentration (Fig. 10). These results
5 indicate that hydrological processes strongly affect the DOC export and thereby the
6 carbon budget in the catchment.

7

8

[Fig. 9]

9

10

[Fig. 10]

11

1 **V. Summary and Outlook**

2 The seasonally concentrated precipitation increases the surface runoff, when the
3 infiltration capacity of the soil decreases during summer monsoon. The outbreak of
4 surface runoff reduced the mean residence time of water in the catchment, and increased
5 DOC export from the surface soil layer. The precipitation also plays an important role in
6 infiltration processes of dissolved material. The precipitation patterns and hydrological
7 processes strongly affect the carbon cycling in the Korean temperate forest during
8 summer monsoon. The increasing occasions of heavy precipitation may not lead to the
9 simultaneous increase of available water resources in the catchment due to the
10 shortening of the water residence time. However, the heavy precipitation will clearly
11 increase material discharge such as DOC. Therefore, the effect of monsoon climate on
12 water and carbon cycling in forest catchment should be critically evaluated on the basis
13 of improved understanding of catchment hydrological and biogeochemical processes.

14 The major interest in HydroKorea and CarboKorea is the generalization of water
15 and carbon flowpaths, storage, residence time, and scaling. Our understandings in water
16 and carbon cycling obtained from the hydro-biogeochemical approaches are limited due
17 to the prescribed spatial scale of the measurements (Moon *et al.*, this issue). The scaling
18 issues are implicitly built into our field measurements and model representations (Kim
19 *et al.*, 2006). The data presented in our study are currently used to calibrate and improve
20 ecohydrological modeling schemes such as the Regional Hydrological and Ecological
21 Simulation System (RHESSys) (see Kim *et al.* this issue), which will be used to
22 estimate long-term and large-scale exchanges of water and carbon in Korean forest
23 catchments.

24

1 **적요**

2 한국 산악 경관에서의 물과 탄소의 순환을 이해하고 예측하기
3 위해서는 물과 탄소의 유역 내 이동 경로, 저류, 체류시간 및 상호작용에
4 대한 정보가 선행되어야 한다. 이와 관련하여 본 논문에서는 HydroKorea 및
5 CarboKorea 연구에서 사용하고 있는 연구 방법들과 현재까지의 주요 결과를
6 소개한다. 유역 내 다양한 수문순환 과정을 이해하기 위해 지하수위 변동,
7 end-member mixing model, 교차상관 분석, 대기 기원의 천연방사성 동위원소를
8 이용하였다. 광릉 산림 유역에서는 지표유출의 기여도가 상대적으로 높았고,
9 강수량과 강수강도 및 패턴의 변화가 물의 유출경로와 체류시간에 영향을
10 주었다. 특히, 몬순으로 인한 강수형태와 유역 내 수문과정의 변화가 탄소
11 순환에 영향을 미쳤는데, 지속적인 강우의 유입이 산림토양의 표층에
12 분포하는 고농도의 용존유기탄소의 유출을 증가시켰다. 본 연구를 통하여
13 시도된 수문순환과정에 대한 정량적인 규명은 기후 변화가 수자원 관리와
14 산림유역 탄소순환에 미치는 영향을 예측하기 위한 과학적 방법론을
15 확립하는데 기여할 것으로 기대된다.

16

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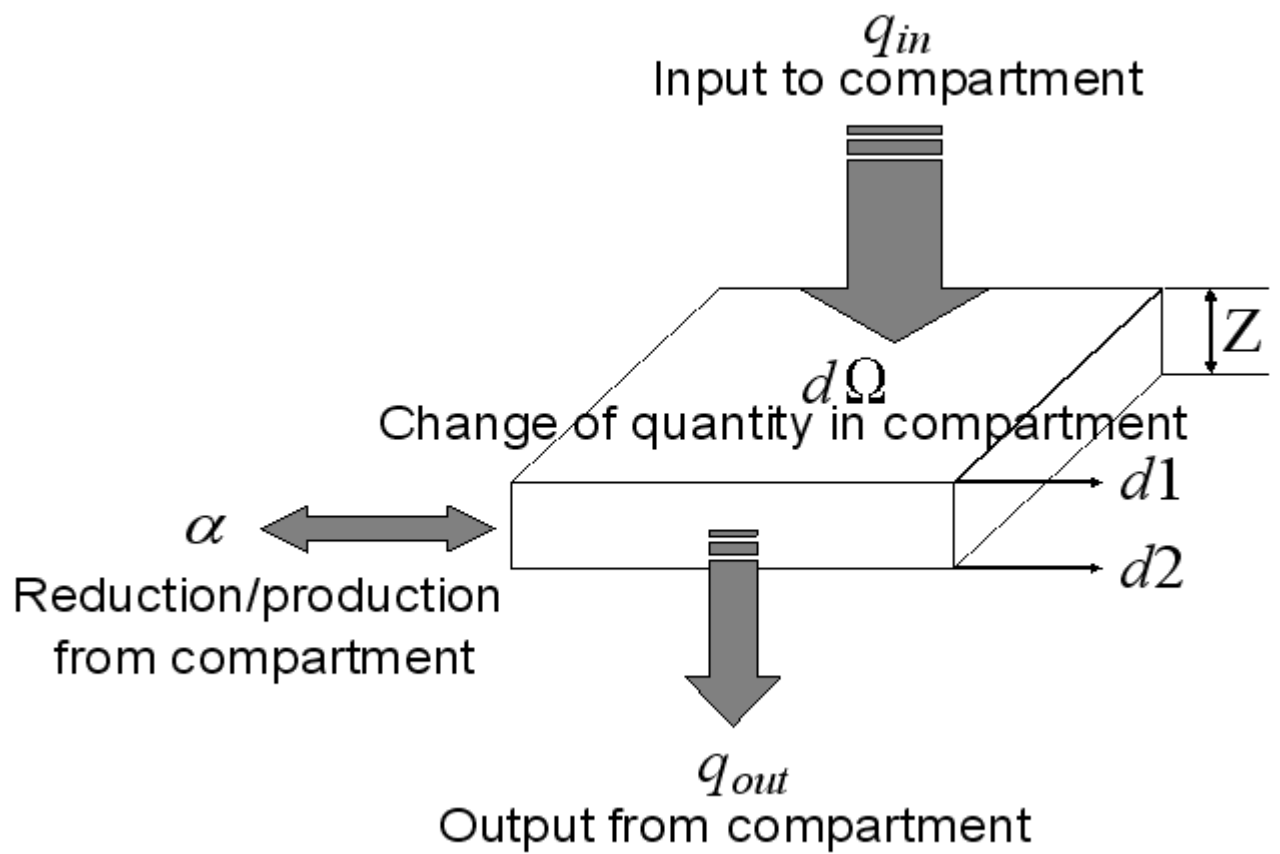
1 Table 1. Global carbon flux to oceans

	Flux (Gt-C yr ⁻¹)	Reference
DOC	0.20	Meybeck 1980
DIC	0.24	Meybeck 1980
POC	0.10	Meybeck 1980
DOC	0.21	Ludwig <i>et al.</i> 1996
POC	0.17	Ludwig <i>et al.</i> 1996
TOC	0.41	Schlesinger and Melack 1981

2

3 DOC : Dissolved Organic Carbon DIC : Dissolved Inorganic Carbon

4 POC : Particulate Organic Carbon TOC : Total Organic Carbon



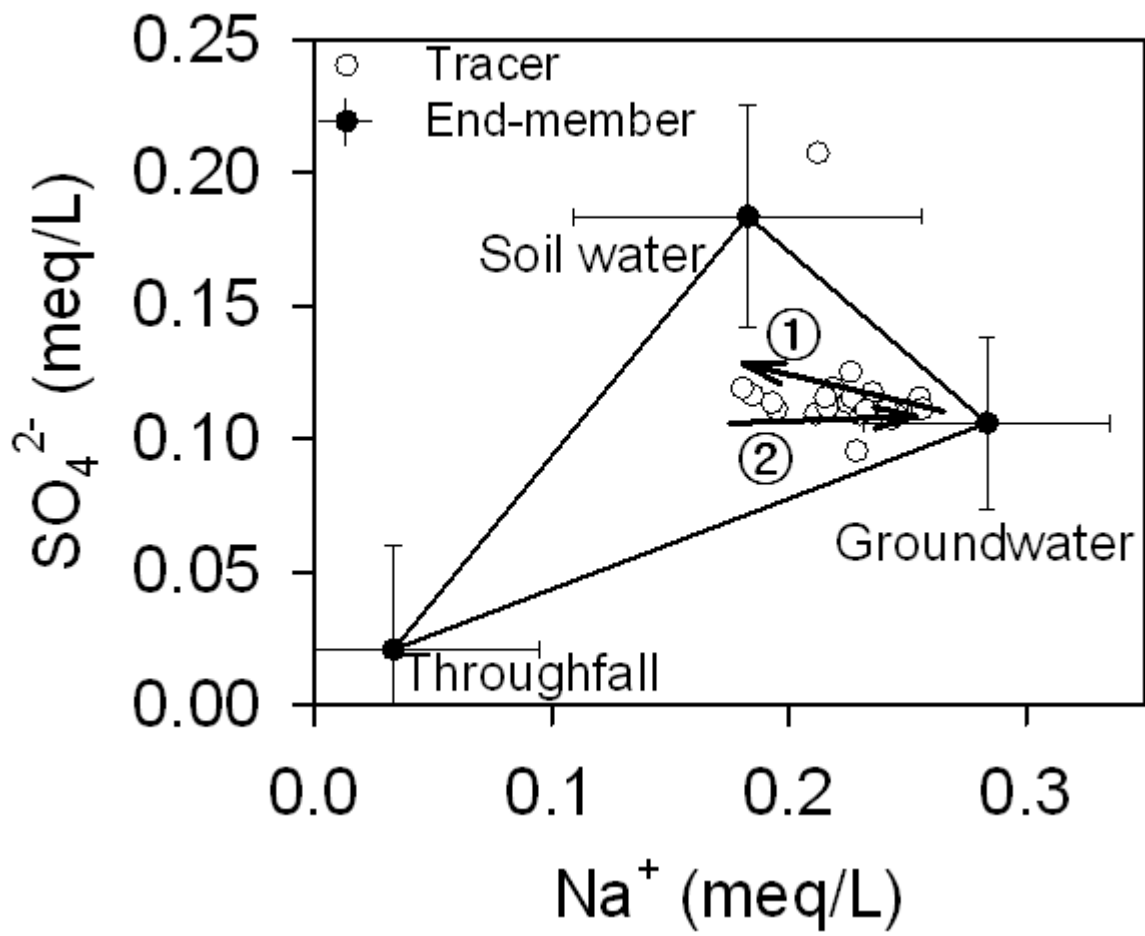
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3 Fig. 1

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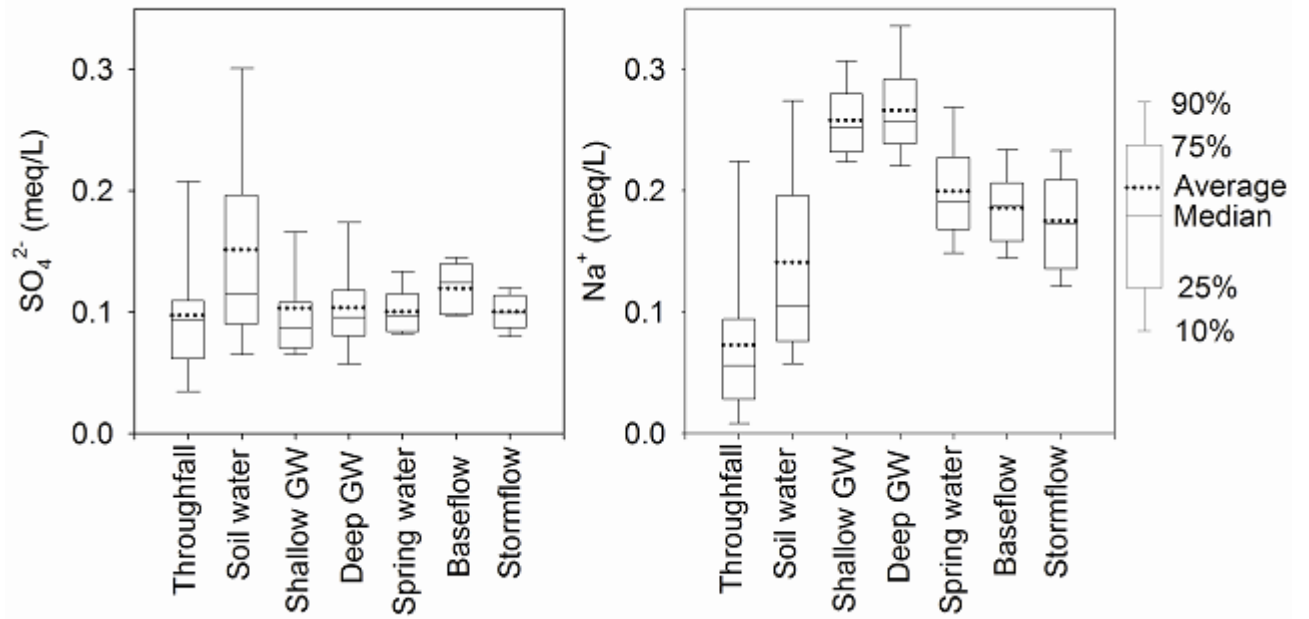
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4 Fig. 2

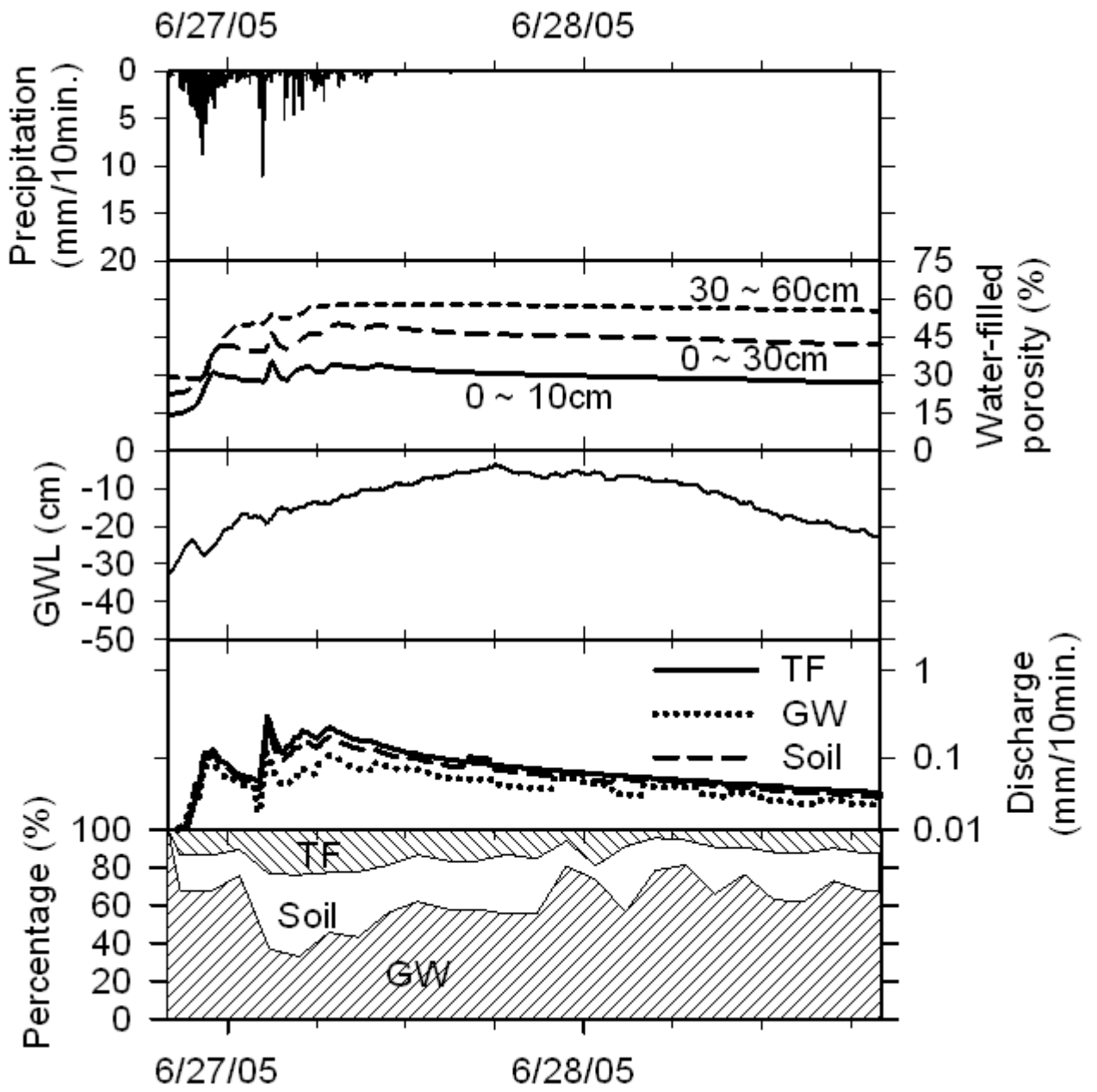
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5 Fig. 3

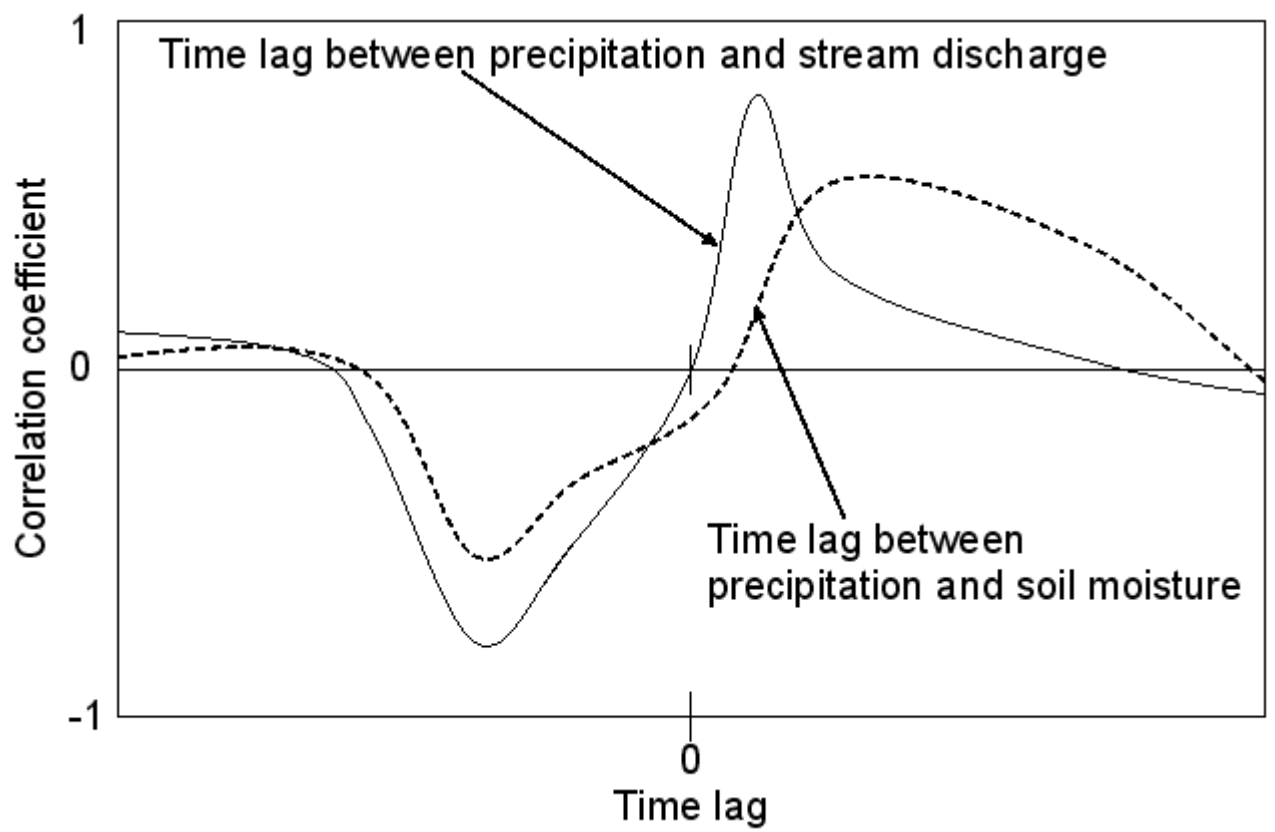


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4 Fig. 4



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4 Fig. 5

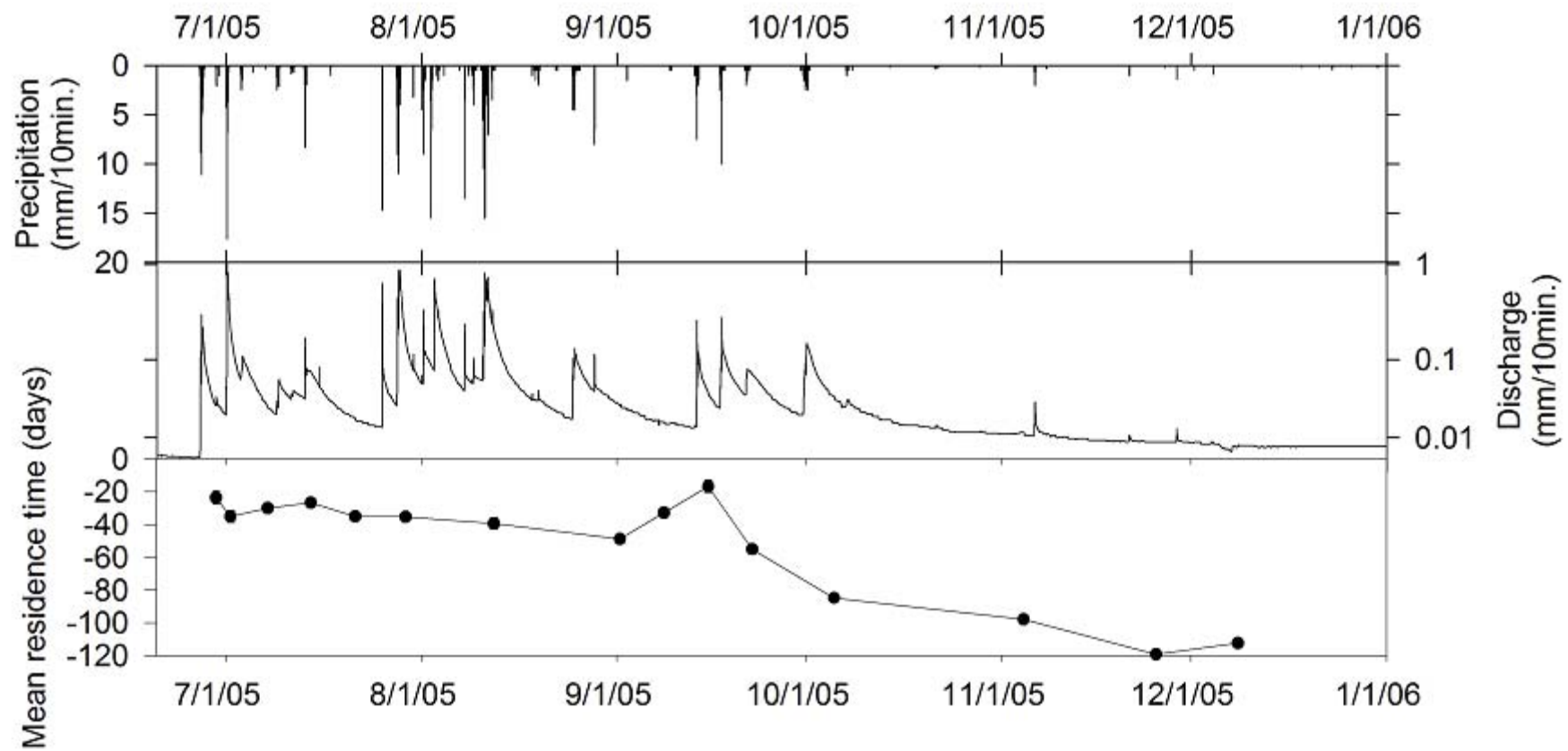
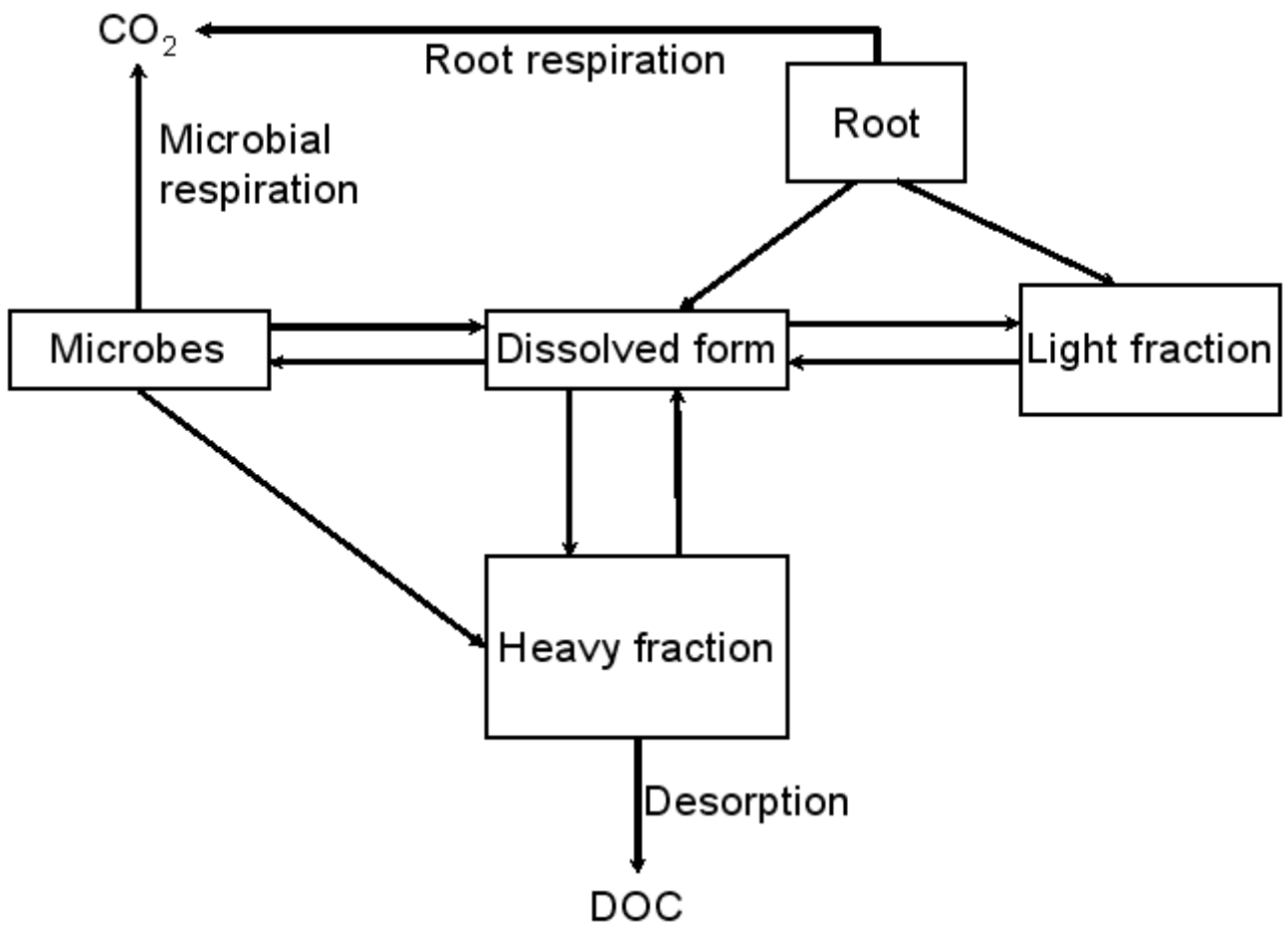
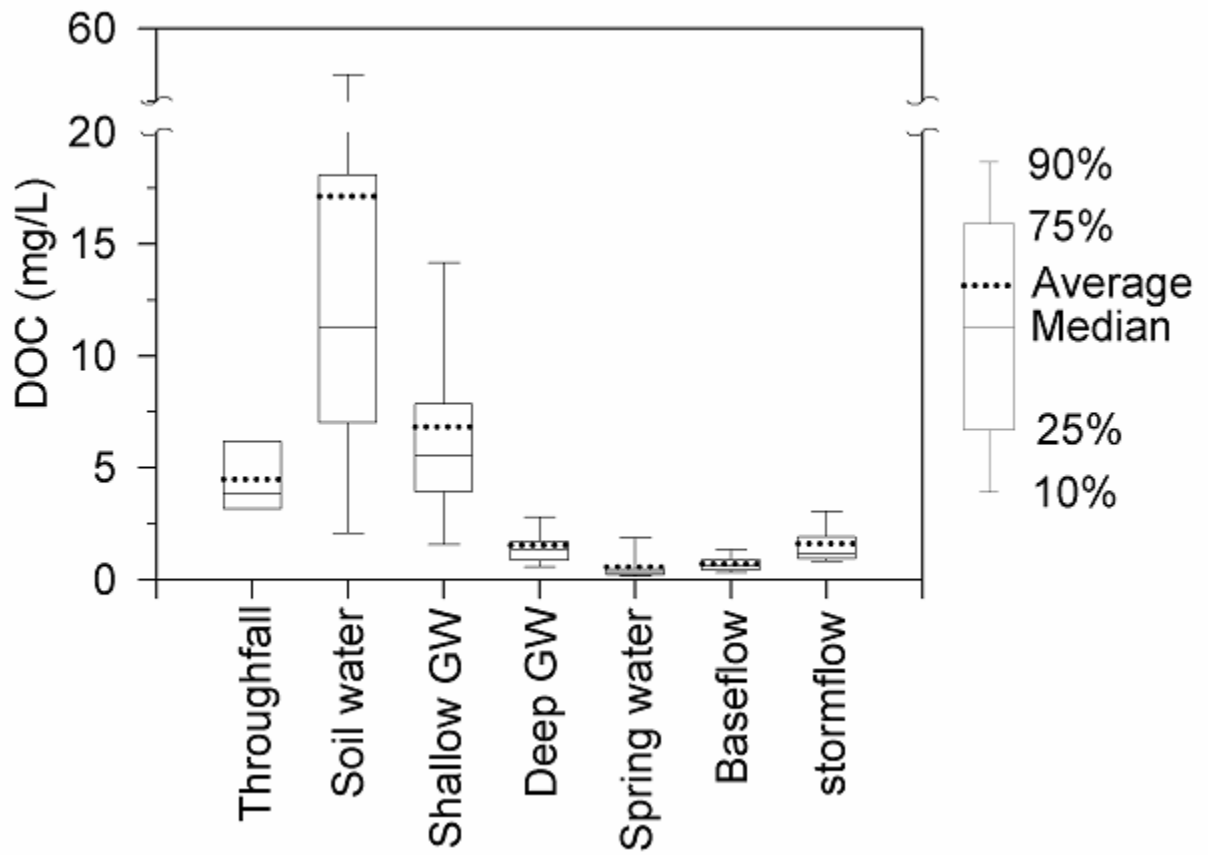


Fig. 6



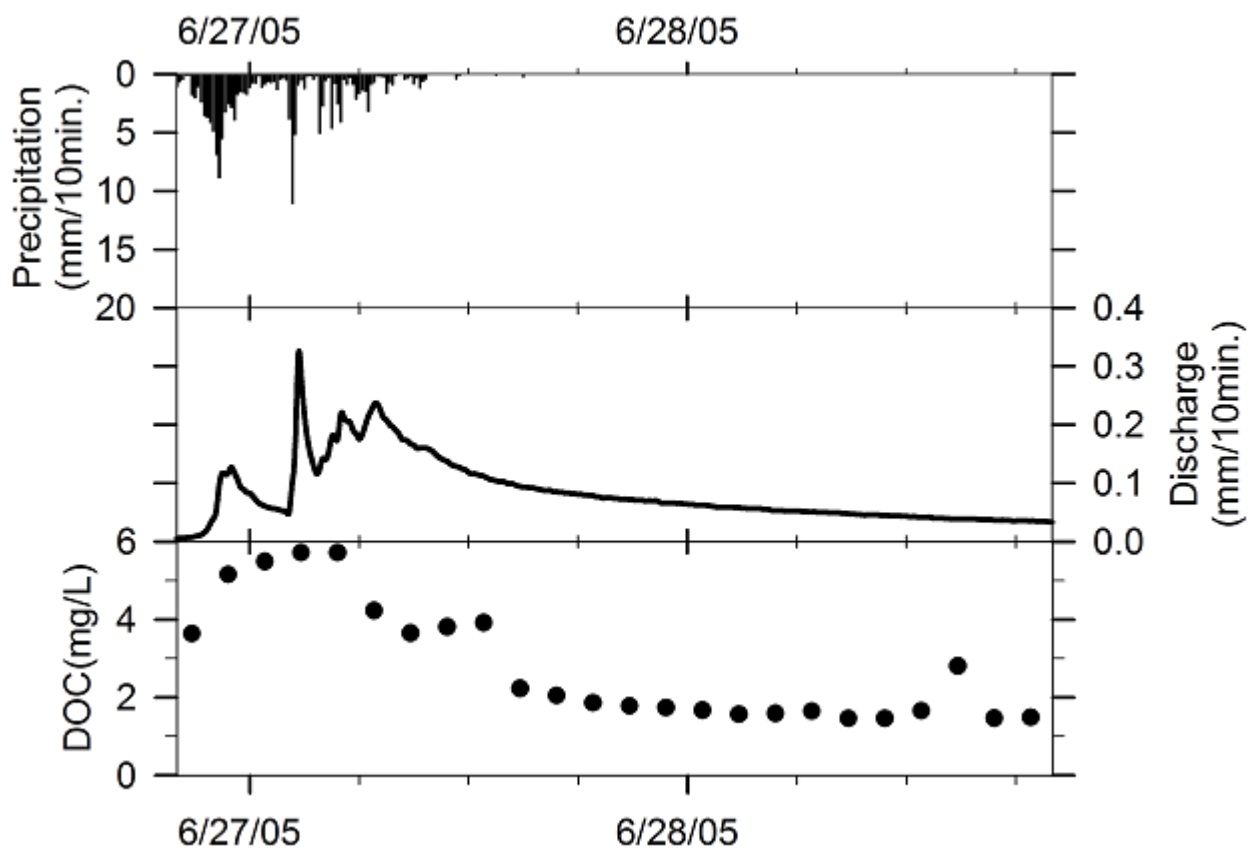
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Fig. 7



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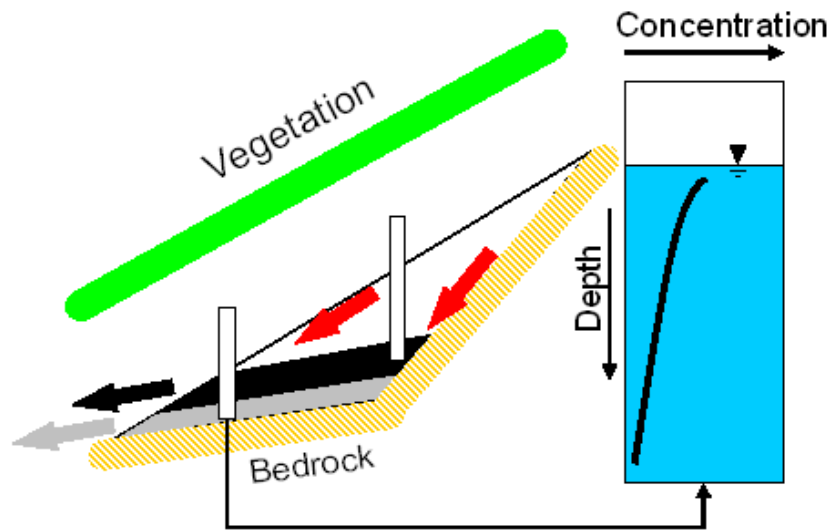
Fig. 8



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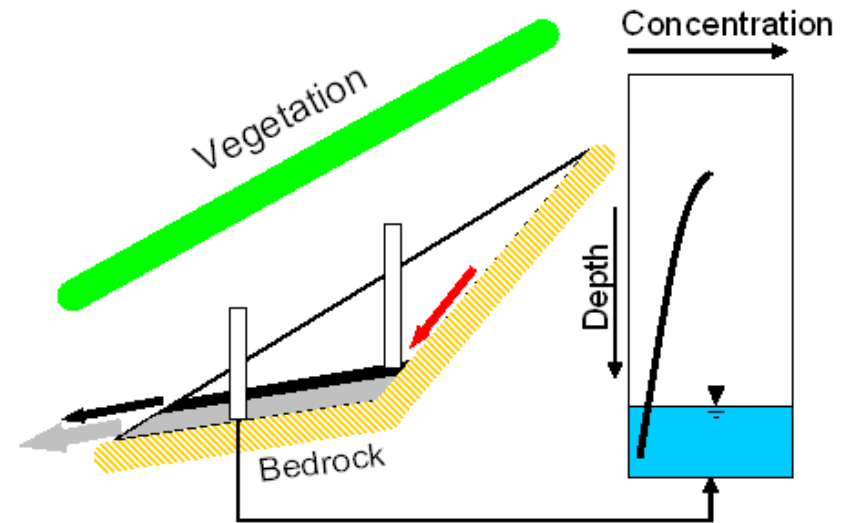
Fig. 9

Storm event period;
Large supply from surface soil



Increase in proportion of surface/subsurface soil discharge
→ Increase in DOC concentration in streamwater

Base flow period;
Small supply from surface soil



Increase in proportion of deep groundwater discharge
→ Decrease in DOC concentration in streamwater

1

2 Fig. 10