Estimation of Bedrock Infiltration on a Weathered Granitic Mountain Covered by Japanese Cypress Forest using Water-Budget and Eddy Covariance Methods

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At a watershed underlain by weathered granitic bedrock (5.99 ha) and covered by Japanese cypress forest, we estimated apparent bedrock infiltration ($Q_B$) by extracting evapotranspiration measured by eddy covariance ($E_{EC}$) from the residual of the water budget ($E_{WB} = \text{precipitation} – \text{runoff}$) over an 8-year period (2001–2008). The degree of interannual variation in $E_{EC}$ was relatively small throughout this period. Conversely, $E_{WB}$ showed larger annual variations and positive dependence on annual precipitation. The difference between $E_{EC}$ and $E_{WB}$ was greater in the years with more precipitation. Therefore, estimated values of $Q_B$ also showed positive dependence on annual precipitation. Observed average annual precipitation, runoff, $E_{EC}$, and estimated $Q_B$ were 1549, 722, 742, and 86 mm year$^{-1}$, respectively. Consequently, $Q_B$ is believed to have amounted to 0–10% (average 5%) of total precipitation. The findings suggest that episodes of rainfall with medium and high intensity induce lower and higher $Q_B$ values, respectively, with similar amounts of precipitation.

1. INTRODUCTION

Recently, the risks of deep-seated landslides have been of concern in Japan due to increases in the frequency of abnormal climate events such as episodes of torrential rain. When deep-seated landslides occur, huge amounts of sediment collapse, with landslides composed of not only shallow soil layers but also the weathered bedrock that originally lay under the shallow soil layers. Although deep-seated landslides are less frequent than shallow landslides, the resulting damage to local communities may be much greater. Many studies have suggested that bedrock groundwater flow plays a significant role in the occurrence of such landslides [e.g., Everett, 1979; Iverson and Major, 1986; Onda et al., 1999; Kato et al., 2000; Montgomery et al., 2002]. Jitousono et al. [2004, 2006] speculated that deep-seated landslides occur due to water accumulating at the boundary zone between a thick, weathered rock layer and an impermeable layer. Thus, quantification of the amount of bedrock infiltration as a bedrock groundwater resource is essential in understanding the causal mechanism behind deep-seated landslides.

The results of several studies have implied the occurrence of water infiltration in permeable bedrock in certain small headwater catchments. Katsura et al. [2009] reported that the hydraulic conductivity of weathered granite bedrock (10$^{-5}$–10$^{-3}$ cm s$^{-1}$) is higher than that of weakly weathered bedrock (10$^{-8}$ cm s$^{-1}$) in deeper layers. Based on intensive observations of the SiO$_2$ concentrations of groundwater and streamwater, Katsuyama et al. [2005] showed that groundwater flow within a permeable weathered granite bedrock contributed to streams year-round in one small headwater catchment. Terajima et al. [1993] examined the water balance in two small catchments (0.087 and 0.255 ha) in a granitic mountain area and suggested that at least 30% and 18% of annual precipitation, respectively, percolates into bedrock. Waichler et al. [2005] suggested that groundwater recharge as a loss term in water balance could account for about 12%
of precipitation in three small catchments (0.06–0.1 ha) in H. J. Andrews Experimental Forest, Oregon, USA. Kosugi et al. [2006] conducted hydrometric observations in different-scale weathered granitic catchments and suggested that annual bedrock infiltration ranged from 35% to 55% of annual precipitation in an unchanneled 0.024-ha headwater catchment underlain by weathered granite. In addition, in a 0.086-ha watershed including an unchanneled 0.024-ha headwater catchment, exfiltration from the bedrock toward the soil layers made up more than half of the annual discharge. Consequently, most studies examining bedrock infiltration in catchments underlain by weathered bedrock suggest the occurrence of remarkable amounts of bedrock infiltration.

Because of the difficulty in making direct field measurements, most studies estimate bedrock infiltration using water-budget methods. Hydrological studies have generally assumed that the fate of precipitation ($P$) is runoff, as measured by a weir ($Q$), change in soil moisture ($\Delta S$), and evapotranspiration as a residual of the water budget ($E_{WB}$). If the amount of bedrock infiltration is not negligible in the water balance, it should be confined to $E_{WB}$ because it cannot be measured by a weir. Therefore, in principle, we can estimate the amount of ‘apparent bedrock infiltration ($Q_B$)’ via extraction of actual evapotranspiration ($E$) from $E_{WB}$, if we assume that $P$ is consumed by $Q$, $\Delta S$, $E$, and $Q_B$. However, in many studies, the unavailability of accurate $E$ data has resulted in uncertainties about the reliability of estimated $Q_B$. Terajima et al. [1993] used the potential evaporation rate (nearly equal to the evaporation from a wet surface) as the evapotranspiration rate and estimated the least amount of $Q_B$. Waichler et al. [2005] used observed values of $E$ from eddy covariance or sap-flux measurements in 1999 and 2000 to simulate changes over a long period (1958–1998). Under the assumption that $Q_B$ is small for a relatively large watershed containing several headwater catchments, Kosugi et al. [2006] applied the evapotranspiration rate estimated by short-term water-budget methods to larger watersheds containing a headwater catchment.

Several methods have been proposed to quantify evapotranspiration independently of other water budget components. The evapotranspiration from each component of an ecosystem, such as transpiration from leaves or evaporation from the soil surface, can be directly measured using a leaf chamber or lysimeter [e.g., Kosugi et al., 2003; Matsumoto et al., 2005; Daikoku et al., 2008; Deguchi et al., 2008]. Also, the evapotranspiration of whole trees can be estimated from sap-flux measurements [e.g., Wilson et al., 2001; Kumagai et al., 2008]. Evaporation of water intercepted by the canopy can be measured by subtracting throughfall and stem-flow water below the canopy from the rainfall above the canopy [e.g., Toba and Ohta, 2005; Deguchi et al., 2006]. However, continuous long-term measurement of these parameters is difficult, particularly when considering spatial variability. For quantification of whole-ecosystem evapotranspiration, micrometeorological techniques have generally been used. As a classic micrometeorological method, gradient (or aerodynamic) and Bowen-ratio (or energy-balance) methods [e.g., Vogt and Jaeger, 1990; Perez et al., 1999] have been used to estimate the turbulent heat fluxes of a warming boundary layer and the vaporisation of water (i.e., sensible and latent heat fluxes, respectively). The evapotranspiration can be deduced from the latent heat fluxes. Such methods estimate energy fluxes by measuring the vertical gradients of air temperature, specific humidity, and wind velocity (the gradient method) between two different heights above the canopy. However, the conditions under which such methods are applicable, such as a vertical logarithmic wind profile (i.e., near-neutral conditions in atmospheric stability), are limited, and obtaining continuous data sets in this way is therefore difficult. Additionally, large errors are common in the application of these methods because in micrometeorology, accurate and precise measurement of small differences among different heights is difficult. In comparison, the eddy covariance method [e.g., Baldocchi et al., 1988; Verma, 1990] can be applied under more variable conditions and with higher reliability by directly measuring rapid fluctuations in air temperature, humidity, and wind velocity at one position above the canopy. Although this method requires measurements with high time resolution, such as at the 1–10 Hz scale, this has recently become possible owing to technological advances such as ultrasonic anemometers and infrared gas analysers. The eddy covariance method is currently recognised as one of the most direct and reliable ways of measuring water, energy, and CO2 fluxes within the boundary layer and is widely used by micrometeorologists worldwide [e.g., Baldocchi et al., 2001].

Despite its advantages, long-term data sets of evapotranspiration collected using the eddy covariance method ($E_{EC}$) are not available for many research sites because it is a relatively recent technique. Thus, few studies have attempted to quantify $Q_B$ using a combination of water budget and eddy covariance methods. Studies comparing $E_{WB}$
and \( E_{EC} \) are also rare, with the exception of some trials that have presented up to 3 years of data [e.g., Wilson et al., 2001; Kosugi and Katsuyama, 2007]. Following Kosugi and Katsuyama [2007] as a precedent for this study, we have monitored evapotranspiration continuously using the eddy covariance method coupled with hydrological observations in a granitic watershed covered by Japanese cypress forest in central Japan (Kiryu Experimental Watershed [KEW], 5.99 ha). The eddy covariance data have been collected for 8 years up to the present. Using these data, we tried to estimate \( Q_B \) based on a combination of the water budget and eddy covariance methods. In addition, we evaluated interannual variations in \( Q_B \) and examined its dependence on annual precipitation taking into account annual rainfall characteristics.

2. MATERIALS AND METHODS

2.1 Site description

Observations were made in the KEW (34°58'N, 136°00'E; Fig. 1a) in Shiga Prefecture, central Japan. The KEW was established in 1967 to study hydrology in the forest and has one of the longest historical hydrological data sets in Japan. The KEW covers 5.99 ha and ranges in elevation from 190 to 255 m (Fig. 1b). The site has a warm-temperate climate; the annual mean air temperature measured at an open screen site from 2001 to 2008 was 13.3°C. The annual mean precipitation from 2001 to 2008 was 1549.4 mm. Rainfall occurs throughout the year, with a peak in summer and little snowfall occurring in winter. The entire watershed is underlain by weathered granite with a high proportion of albite. The A0 horizon has a humus form with no apparent H horizon. Ohte et al. [1995, 1997] detailed the physical and chemical properties of the soil in the watershed. The KEW is dominated by Chamaecyparis obtusa Sieb. et Zucc. (Japanese cypress, an evergreen conifer) planted in 1959. Japanese red pine and several species of deciduous broad-leaved trees are sparsely present. Eurya japonica Thunb. dominates the forest floor. In 2005, the stand density of trees greater than 5 cm diameter at breast height was 1754 trees ha\(^{-1}\), with an average height of 16.6 m, maximum height of 24.0 m, and basal area of 49.4 m\(^2\) ha\(^{-1}\). The total leaf area index (LAI), measured using an LAI-2000 plant canopy analyser (LI-COR, USA), ranged from 4.5 to 5.5, with small seasonal variation.

2.2 Estimation of apparent water supply to deeper layers

To estimate the amount of bedrock infiltration, we examined the water budget at KEW according to the water-balance equation as follows:

\[
P = Q + E_{WB} + \Delta S, \quad (1)
\]

where \( P \) is precipitation (mm), \( Q \) is the runoff rate (mm), \( E_{WB} \) is the residual (mm), which has been regarded as the evapotranspiration rate in most hydrological studies, and \( \Delta S \) is the change in water storage within the soil layer (mm). \( \Delta S \) was assumed to be nearly zero on a water-year basis, which we defined as one calendar year (i.e., from 1 January through 31 December) because winter is the driest season in this region. In Eq. (1), if bedrock infiltration exists at KEW, it should be included in
$E_{WB}$ as follows:

$$E_{WB} = E + Q_b,$$

where $Q_b$ is apparent bedrock infiltration (mm), and $E$ is actual evapotranspiration (mm). In this study, we measured the evapotranspiration rate using the eddy covariance method ($E_{EC}$, mm) and estimated $Q_b$ from the balance of $E_{WB}$ and $E_{EC}$ using Eq. (2). Details of the methods used to obtain $P$ and $Q$, and $E_{EC}$, are given in Sections 2.3 and 2.5, respectively.

### 2.3 Hydrological observations

Precipitation ($P$, mm) was measured using a tipping bucket (capacity 0.5 mm of water) and corrected using another storage-type rain gauge at the meteorological station. The runoff rate from the entire experimental watershed (5.99 ha) has been measured at a 90° V-notch weir on the stream (see Fig. 1a and 1d) since 1967 [Fukushima, 1988]. Stage height was recorded continuously, and $Q$ was calculated using the measured relationship between stage height and the actual discharge rate from the weir. A reference stage height was checked every 2 weeks using a hook gauge. Continuous hydrological observations have been made here since 1972, and data from 8 years (2001–2008) were used in this study.

### 2.4 Micrometeorological and energy balance observation

The energy balance above a forest canopy can be described as follows:

$$R_n - G - J = H + \lambda E,$$

where $R_n$ is the net radiation (W m⁻²), $G$ is the ground heat flow (W m⁻²), $J$ is the change in heat storage in the canopy layer (W m⁻²), $H$ is the sensible heat flux (W m⁻²), and $\lambda E$ is the latent heat flux (W m⁻²). The right-hand side of Eq. (3) represents the turbulent heat fluxes, and the left-hand side expresses the available energy for turbulent heat fluxes. At the KEW, we have been measuring or estimating each energy-balance component since 2001. The turbulent fluxes above the canopy were measured using an eddy covariance technique (see Section 2.5), missing $\lambda E$ or $H$ data were replaced by estimation using Eq. (3).

$R_n$ values were calculated by downward and upward short-wave and long-wave radiation measured at a height of 28.5 m on the micrometeorological observation tower (Fig. 1b). Ground heat flow was measured by three heat-flux plates installed at a depth of 0.01 m below the forest floor within a 10-m circle centred on the tower, and the average value of the three sensors was used as $G$. $J$ was estimated by the method proposed by Moore and Fisch [1986], with measurements of vertical profiles of air temperature and relative humidity (29, 20, 10, 4 and 1 m heights). Details of these measurements and estimations are given in Kosugi et al. [2007].

### 2.5. Measurement of evapotranspiration by the eddy covariance method

In the eddy covariance method, the basic equations for sensible heat flux ($H$, W m⁻²) and latent heat flux ($\lambda E$, W m⁻²) can be written as follows:

$$H = \rho C_p \overline{w'T'},$$

$$\lambda E = \lambda \rho w' q',$$

$$w' = w - \overline{w},$$

$$T' = T - \overline{T},$$

$$q' = q - \overline{q},$$

where $\rho$ is the density of air (kg m⁻³), $C_p$ is the specific heat of air at constant pressure (J kg⁻¹ K⁻¹), $\lambda$ is the latent heat of vaporisation of water (J kg⁻¹), and $w$, $T$, and $q$ are the vertical components of wind speed (m s⁻¹), air temperature (°C), and specific humidity (kg kg⁻¹), respectively. The primes indicate the instantaneous deviation from the mean, and overbars denote temporal averages.

In the KEW, those fluxes are measured at a height of 28.5 m on the micrometeorological observation tower using an infrared-red gas analyser (IRGA) for $q$ and three-dimensional ultrasonic thermo-anemometer for $w$ and $T$. The signals were sampled at a frequency of 10 Hz, and calculations of fluxes were made every 30 min. Systems before and after 30 April 2002 are closed- and open-path systems, respectively. In the closed-path system, air sampled at the same position as the anemometer was directed to IRGA by a tube, whereas in the open-path system, an open-type IRGA sensor was placed at the same position as the anemometer. Details of each system and corrections for the gained data are given in Kosugi et al. [2007].

At many study sites where researchers have used the eddy covariance method, smaller turbulent heat fluxes ($= H + \lambda E$) than the available energy ($= R_n - G - J$) have been reported [e.g., Turnipseed et al., 2002; Wilson et al., 2002]. Such an imbalance has also been shown in the KEW, with observed turbulent heat fluxes about 90% of the available energy [Kosugi et al., 2007]. To overcome this problem, we carried out an energy-budget correction. For daytime data, to address the energy balance, we inflated both $H$ and $\lambda E$ with the same Bowen ratio ($= H/\lambda E$). For nighttime data, we inflated only $H$ because nighttime evapotranspiration is thought to be negligible. If $H$ or $\lambda E$ data were not available, we based daytime inflation on the assumption that it was...
underestimated by 0.9%, and we did not inflate it for the nighttime [Kosugi et al., 2007].

Linear trends in $T$ and $q$ were removed. Data spikes with standard deviation values exceeding 3.5 were replaced with interpolated values. Data points that were outside a set range (e.g., caused by analogue communication errors) were replaced. Spikes or out-of-range data exceeding 1% of the total data were rejected for quality reasons. Data that did not meet the stationarity criteria [Foken and Wichura, 1996; Aubinet et al., 2000] were also rejected. Rainfall events tend to cause large errors in measurement of $q$, especially for the closed-path system. Therefore, $\lambda E$ data obtained during and after episodes of rainfall (from 30 min before to 24 h after a rainfall event) were rejected and replaced by $\lambda E$ estimated from the energy balance ($= R_n - G - J - H$). Gaps were also filled using the energy-balance equation if data for only one of $\lambda E$ or $H$ were missing. Several studies have reported that during rainfall, $H$ can be measured by a sonic anemometer [e.g., Mizutani et al., 1997; van der Tol et al., 2003]. In the KEW, during and after rainfall events, almost the same variations in $\lambda E$ have been simulated using the calibrated multi-layer model and estimated by the energy budget with eddy covariance $H$ [Takanashi et al., 2003]. Therefore, in this study, we used $H$ measured by a sonic anemometer regardless of the weather conditions.

When both $\lambda E$ and $H$ were not available, we filled gaps using the following three methods.

A) For days with more than 80% of half-hourly $\lambda E$ data, we calculated daily $E_{EC}$ from $\lambda E$, and for each year, we obtained the relationship between daily $E_{EC}$ and available energy ($= R_n - G - S$) for fine days and rainy days, respectively. Then, we filled gaps in daily $E_{EC}$ data from available energy using obtained relationships to calculate the annual $E_{EC}$.

B) Based on half-hourly data, we obtained the relationship between $\lambda E$ and available energy for each year and filled gaps in $\lambda E$ data using this relationship. Then, we calculated the annual $E_{EC}$ from gap-filled half-hourly $\lambda E$.

C) Based on half-hourly data, for each year, we obtained the relationship between $\lambda E$ and available energy for dry and wet periods during and after 6 h of a rainfall event, respectively. Then, we filled gaps in $\lambda E$ data using this relationship to calculate the annual $E_{EC}$.

Finally, we calculated the annual amount of evapotranspiration ($E_{EC}$, mm year$^{-1}$) using our obtained daily or half-hourly data sets of $\lambda E$. We believed that the last method is more accurate than the first two. Method A uses daily data and therefore cannot account for variation within a day. Evapotranspiration during wet periods is generally larger than that in dry periods under given environmental conditions due to evaporation intercepted by the canopy. Because Method A cannot account for the difference in the length of dry and wet periods within a day, errors will become large, especially for data obtained during rainy days. Method B is based on data taken every half hour, but it cannot account for the differences within the dry and wet periods. Method-C, however, addresses the problems with Methods A and B and is therefore preferable.

It is possible that disagreement between the measurement area of eddy covariance data and the watershed area might affect the results. However, from the tower, the forest area reaches approximately 750 m in a northwest direction and more than 2000 m in the other directions. Takanashi et al. (2005) reported that for 92% of the data, more than 80% of the measured flux could be expected to come from within the Japanese cypress forest in or

<table>
<thead>
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<th>Year</th>
<th>$P$ (mm year$^{-1}$)</th>
<th>$Q$ (mm year$^{-1}$)</th>
<th>$E_{WB}$ (mm year$^{-1}$)</th>
<th>$E_{EC_A}$ (mm year$^{-1}$)</th>
<th>$E_{EC_B}$ (mm year$^{-1}$)</th>
<th>$E_{EC_C}$ (mm year$^{-1}$)</th>
<th>$A_E$ (MJ m$^{-2}$ day$^{-1}$)</th>
<th>$R_n - G - J$ (MJ m$^{-2}$ day$^{-1}$)</th>
<th>$Q_{B_A}$ (mm year$^{-1}$)</th>
<th>$Q_{B_B}$ (mm year$^{-1}$)</th>
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around the KEW during the daytime. Therefore, we believe that we have obtained representative evapotranspiration values for the Japanese cypress forest at the KEW by using eddy covariance measurements.

3. RESULTS

3.1 Runoff, precipitation, and residual in the water balance

Table 1 shows the observed annual values of water budget components at the KEW. From 2001 to 2008, no data were missing for runoff. The average and standard deviation of annual precipitation (P), runoff (Q) and residual (E\(_{\text{WB}} = P - Q\)) from 2001 to 2008 were 1549 ± 312, 722 ± 260, and 827 ± 69 mm year\(^{-1}\), respectively. Q and E\(_{\text{WB}}\) accounted for 45 ± 8% and 55 ± 8% of P, respectively.

3.2 Evapotranspiration by eddy covariance

The average (± standard deviation) data-acquisition percentage for the half-hourly evapotranspiration rate measured by the eddy covariance method (\(E_{\text{EC}}\)) was 85 ± 9% from 2001 to 2008 (Table 1). Fig. 2 shows long-term variations in daily \(E_{\text{EC}}\) measurements (mm day\(^{-1}\)). For each year, daily \(E_{\text{EC}}\) data showed similar seasonal variations. In the summer, \(E_{\text{EC}}\) under dry canopy conditions reached about 4 mm day\(^{-1}\). As shown in Fig. 3a, variations in \(E_{\text{EC}}\) were significantly dependent on available energy. Because of evaporation from the wet canopy during and after rainfall, \(E_{\text{EC}}\) was larger for rainy days than for fine days. On an annual basis, variations in \(E_{\text{EC}}\) were small, and no dependence on available energy was detected (Table 1 and Fig. 3b). Annual \(E_{\text{EC}}\) values were not largely different when gap-filling methods A–C were used. The maximum difference among the gap-filling methods was 28 mm year\(^{-1}\) in 2001, the year with the lowest data acquisition (69%, Table 1). The average and standard deviation of annual \(E_{\text{ECs}}\) obtained using gap-filling methods A–C were 745 ± 15, 741 ± 21, and 742 ± 21 mm year\(^{-1}\), respectively.
3.3 Apparent bedrock infiltration and dependence on precipitation

Fig. 4 (a and b) shows the interannual relationships between \( P \) and \( E_{WB} \) and between \( P \) and \( E_{EC} \), respectively. Whereas annual variations in \( E_{WB} \) were relatively large and positively correlated with annual variations in \( P \), the variations in \( E_{EC} \) were small and not dependent on \( P \). The values of \( E_{EC} \) ranged between 715 and 780 mm year\(^{-1} \), whereas the \( E_{WB} \) values ranged from 739 to 940 mm year\(^{-1} \). If we assume that \( E_{EC} \) was the actual evapotranspiration rate (\( E \)), some residual can be obtained from the balance of \( E_{WB} \) and \( E_{EC} \). In this study, we defined this unknown component as apparent bedrock infiltration (\( Q_B = E_{WB} - E_{EC}, \) mm year\(^{-1} \)) and calculated the annual \( Q_B \) from the data for 2001–2008. These results gave \( Q_B \) generally ranging between 0 and 200 mm year\(^{-1} \) (average \( 83 \pm 68, 87 \pm 61, \) and \( 86 \pm 64 \) mm year\(^{-1} \) for gap-filling methods A, B, and C, respectively; Table 1 and Fig. 5a). Table 2 shows the percentage consumption of precipitation by \( Q \), \( E_{EC} \), and \( Q_B \). The \( Q_B \) consumption percentage ranged from 0% to 10% of the precipitation. Average observed runoff (\( Q \)), \( E_{EC} \) and estimated \( Q_B \) accounted for 45 ± 8%, 50 ± 10%, and 5 ± 3–4% of the total precipitation (\( P \)) through the 2001–2008 period, respectively (clear differences between the results of different gap-filling methods were not detected). \( Q_B \) showed large annual variation and a strong positive correlation with annual total precipitation (\( P \); Fig. 5a). The sensitivity of \( Q_B \) to \( P \) was smaller than that of runoff (\( Q \); Fig. 5b). In addition, the correlation between \( Q_B \) and \( P \) (|\( r | = 0.78–0.81) was weaker than that between \( Q \) and \( P \) (|\( r | = 0.99).

### Table 2

Percentages (%) of runoff (\( Q, \) mm year\(^{-1} \)), evapotranspiration by the eddy covariance method (\( E_{EC}, \) mm year\(^{-1} \)), and apparent bedrock infiltration (\( Q_B, \) mm year\(^{-1} \)) of precipitation (\( P, \) mm year\(^{-1} \)) from 2001 to 2008. Ranges for percentages of \( E_{EC} \) and \( Q_B \) express the ranges of values among gap-filling methods A, B, and C.

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<th>( E_{EC} )</th>
<th>( Q_B )</th>
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<tr>
<td>2008</td>
<td>47</td>
<td>45–46</td>
<td>7–8</td>
</tr>
<tr>
<td>Average</td>
<td>45</td>
<td>50</td>
<td>5</td>
</tr>
</tbody>
</table>

### Fig. 5

(a) The relationship between the annual apparent bedrock infiltration (\( Q_B, \) mm year\(^{-1} \)) and annual total precipitation (\( P, \) mm year\(^{-1} \)). Red, yellow, and open dots show \( Q_B \) with gap-filling methods A, B, and C, respectively. (b) The relationship between the annual runoff (\( Q, \) mm year\(^{-1} \)) and annual total precipitation (\( P, \) mm year\(^{-1} \)). Fitting lines for \( Q_B \) made using gap-filling method C are shown by a dotted line.

### 4. DISCUSSION

Several studies in several catchments have estimated \( Q_B \) based on the water budget. However, few studies have estimated \( Q_B \) over long periods using simultaneously and directly measured evapotranspiration data as well as runoff and precipitation data. In this study of a watershed underlain by granitic weathered bedrock (5.99 ha) and covered by Japanese cypress forest, we estimated bedrock infiltration over an interval of 8 years (2001–2008) by examining the water budget using the eddy covariance method, the most reliable means of canopy-scale evapotranspiration measurement. At the KEW, Kosugi and Katsuyama [2007], who did not assume that bedrock infiltration occurred in their water budget, concluded that estimated evapotranspiration (i.e., \( E_{WB} \)) was roughly in accord with that measured by the eddy covariance method (i.e., \( E_{EC} \)) using 3 years of eddy covariance data (2001–2003). However, using the 8 year data set (2001–2008) in this study, the annual \( E_{WB} \) tended to be larger than \( E_{EC} \) for many years, especially for the years with high amounts of precipitation. The difference between \( E_{WB} \) and \( E_{EC} \) was sometimes more than 100 mm year\(^{-1} \). A large portion of this term is thought to represent bedrock infiltration, although this could possibly be accounted for by measurement error or by some other unknown factors.
output that cannot be measured by eddy covariance or weir (e.g., evaporation of small droplets produced by raindrop impacts on plant canopies; Dunin et al., 1988; Murakami, 2006). Hence, we refer to this term as “apparent bedrock infiltration ($Q_B$”).

To examine the difference in annual $E_{EC}$ (and thus $Q_B$) resulting from using gap-filling methods, we calculated the annual $E_{EC}$ and $Q_B$ using three different gap-filling methods (see Section 2.5). Annual $E_{EC}$ values were shown not to significantly differ among the various gap-filling methods. The maximum difference among gap-filling methods was 28 mm year$^{-1}$ in 2001, the year in which data acquisition was lowest (69%; Table 1). Therefore, for the years with the largest gaps in data, meticulous methods are necessary for gap filling. For gap filling, this study obtained the relationship between $\lambda E$ and available energy for each year. However, to fill the gaps more accurately, we should use a relationship based on data sets with higher temporal resolution (e.g., month or season).

In this study, $Q_B$ was estimated to consume 0–10% (average 5%) of the precipitation (Table 2). This value is smaller than that shown in other studies [more than 10% of precipitation; e.g., Terajima et al., 1993; Waichler et al., 2005; Kosugi et al., 2006]. Those studies all focused on headwater catchments and catchments that were much smaller in area (0.02–0.26 ha) than the KEW (5.99 ha). Many studies have indicated an important contribution of bedrock groundwater to storm-runoff generation [e.g., Terajima and Moroto, 1990; Wilson et al., 1993] and to base-flow discharge [e.g., Mulholland, 1993; Burns et al., 1998]. The KEW is a watershed that includes several headwater catchments. Kosugi et al. (2006) reported that $Q_B$ ranged from 35% to 55% of the precipitation in a particular headwater catchment within the KEW. Therefore, a large portion of bedrock infiltration in this headwater catchment was thought to contribute to runoff downstream. Note that a non-negligible amount of $Q_B$ still exists in a relatively larger watershed. Until now, almost hydrologists have included bedrock infiltration within their water budgets. Thus, the results obtained in this study will provide valuable suggestions for future hydrological studies including those examining the risk of deep-seated landslides.

In the KEW, turbulent heat fluxes obtained using the eddy covariance method accounted for only about 90% of the available energy [Kosugi et al., 2007]. This imbalance problem is common at many observation sites where the eddy covariance method has been used [e.g., Wilson et al., 2002]. To correct this imbalance in energy balance, we inflated turbulent heat fluxes (see Section 2.5). However, because the reasons for the imbalance have not been determined, the adequacy of the inflation correction of turbulent heat fluxes is also unclear. If we had not corrected the energy balance, estimated $Q_B$ values would have become larger.

Episodes of heavy rainfall can be a direct trigger of deep-seated landslides [e.g., Jitousono et al., 2008]. Heavy rainfall may result in large water accumulation in the boundary zone between a layer of thick weathered rock and an impermeable layer, and this has been thought to cause deep-seated landslides [e.g., Jitousono et al., 2004, 2006]. The results obtained in this study suggest that bedrock infiltration increases according to the amount of precipitation (Fig. 5a) and that rainfall is indeed a sustaining part of the causal mechanism behind deep-seated landslides. From Figure 5b, we see that a larger amount of precipitation was consumed as runoff than as bedrock infiltration. Therefore, the KEW was thought to be a watershed in which the risk of erosion or shallow landslides was relatively higher than the risk of deep-seated landslides. However, the amount of bedrock infiltration considered dangerous in terms of the risk of deep-seated landslide has not yet been clarified. Therefore, we cannot properly discuss the role of infiltration in the generation of deep-seated landslides at the KEW. Further studies relating bedrock infiltration to structure under the soil layer are needed.

Significant differences in $Q_B$ were observed even when the ranges of $P$ were very similar (i.e., 2001 vs. 2007, 2004 vs. 2008, and 2003 vs. 2006; Fig. 5a). To examine the factors causing these fluctuations, we looked at the characteristics of rainfall in those years. Fig. 6 (a–d) shows the descending order of each half-hourly precipitation amount (mm 30 min$^{-1}$) in each year of 2001–2008, 2001 vs. 2007, 2004 vs. 2008, and 2003 vs. 2006, respectively. In 2001 and 2004, which had smaller $Q_B$ values than the reference years (2007 and 2008, respectively), rainfall episodes of medium intensity (5–10 mm half-hour$^{-1}$) were more frequent than in the reference year. In 2007 and 2006, which had larger $Q_B$ values than the reference years (2001 and 2003, respectively), high-intensity rainfall events with more than 20 mm half-hour$^{-1}$ occurred. Such rainfall events did not occur in the reference years. Although such high-intensity rainfall events also occurred in 2004, medium-intensity rainfall events in 2004 were markedly more pronounced than in any of the other years (Fig. 6a). These findings suggest that medium- and high-intensity rainfall events induce smaller and larger $Q_B$ values, respectively, for similar total amounts of precipitation.
Fig. 6 (a) The descending order of precipitation (mm) per 30 min for each year; (b) for 2001 and 2007; (c) for 2004 and 2008; (d) for 2003 and 2006.

A combination of the water budget and eddy covariance methods is thought to be valid for reliable estimation of $Q_B$. However, as mentioned above, several assumptions are still made in the eddy covariance method. To estimate $Q_B$ with greater reliability, cross-checks of evapotranspiration determined by using several different measurement methods should be conducted. To estimate $Q_B$, we examined its dependence on rainfall characteristics on an annual basis. However, the properties of each rainfall event (e.g., amount, intensity, and duration) were thought to more directly and accurately characterise the amount of bedrock infiltration as a causal factor of deep-seated landslides. Knowledge obtained from these trials should contribute to future advances in our understanding of hydrological processes and elucidate the causal mechanisms of deep-seated landslides.

5. CONCLUSIONS

To estimate bedrock infiltration, for 8 years (2001–2008) we measured the main components of the water budget (i.e., precipitation, runoff, and evapotranspiration) at a watershed underlain by granitic weathered bedrock (5.99 ha) and covered by Japanese cypress forest. As a reliable and direct method of measuring evapotranspiration, we used the eddy covariance method. Results indicated that observed evapotranspiration showed low interannual variation. In contrast, the residual of the water budget (= precipitation – runoff), which has been regarded as evapotranspiration in most hydrological research, showed larger interannual variation and larger values than the observed values of about 0–200 mm year$^{-1}$. Under the assumption that no other unknown water budget components existed, this difference between the observed evapotranspiration and the residual was believed to indicate bedrock infiltration. Average runoff, evapotranspiration, and estimated bedrock infiltration constituted 45%, 50%, and 5% of total the precipitation, respectively, through the period. The estimated annual bedrock infiltration showed a positive dependence on annual precipitation. In addition, rainfall of medium and high intensity tended to induce smaller and larger bedrock infiltration, respectively, under similar amounts of precipitation. As the eddy covariance method is valid for continuous long-term measurement of evapotranspiration, the combination of water budget and eddy covariance methods will be an effective way to clarify the mechanism underlying the generation of deep-seated landslides as well as various hydrological processes at mountainous watersheds.

ACKNOWLEDGEMENTS: We thank the pioneers and current colleagues who have been conducting hydrometeorological observations and research at KEW. The hydrological research at KEW was initiated by Prof. Yoshihiro Fukushima (Tottori University of Environmental Studies) and Prof. Masakazu Suzuki (The University of Tokyo); the management of the hydrological instruments and database were continued by Dr. Nobuhito Ohte, Dr. Ken’ichirou Kosugi, Mr. Yuji Kominami, Dr. Akihiro Nakamura, Dr. Katsunori Tanaka, Dr. Hiroki Tanaka, and colleagues at the laboratories of Forest Hydrology and Erosion Control at Kyoto University. Eddy covariance and meteorological observations were conducted with the assistance of Dr. Hiroki Tanaka, Mr. Masato Yano, Mr. Tatsuya Katayama, Mr. Takumi Wada, Mr. Tomonori Mitani, Mr. Yusuke Fukui, Ms. Rie Fukui, Mr. Naoto Yokoyama, Ms. Chika Soda, Mr. Takuya Matsumoto, Mr. Ryoji
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