Chapter V Discussion

V-1: Relationship between drainage area and discharge

Relationship between drainage area and discharge was analyzed with the spatial data on stream flow. The previous models for channel initiation have assumed that discharge increases in proportion to drainage area (Montgomery and Dietrich, 1989; Dietrich *et al.*, 1993). This section discusses whether or not this assumption is applicable to the investigated area.

Discharge at each first-order stream site and spring site was plotted against drainage area in two cases of base-flow condition (Figure 26) and two cases of storm-flow condition (Figure 27). Drainage area of each observation site is shown in Table 3. Figure 26 shows scattered distribution of area-discharge plots in the base-flow condition. Although time required for measurements (2.0 – 3.7 hours) would produce a statistical error in area-discharge plots, the data show that discharge clearly increases with increasing drainage area in the storm-flow condition (Figure 27). Discharge data at first-order stream sites show more scattered plots for both runoff conditions than those at spring sites. This scattered distribution at stream sites may be caused by water infiltration into bedrock fractures.

Relationship between drainage area, A, and spring discharge, Q_s, was statistically analyzed with simple least squares linear regression. Spring discharge rather than stream discharge should affect channel initiation; hence, only the spring discharge was used for the analysis. The regression lines, which are applicable for both of the base-flow and storm-flow conditions, are expressed as follows:

$$Q_s = \alpha_s (A - \beta_s) \tag{5}$$

where A is drainage area in m^2, Q_s is spring discharge in $m^3 s^{-1}$, α_s and β_s are the coefficients obtained from the regression analysis. Here, the linear regression with an intercept was used because of the following reasons: (1) regression analysis with a power function (log-transformed data) is less significant, because the range of drainage area is small, (2) the regression line is not necessary to pass through the origin (i.e. $Q_s = 0$ and $A = 0$), because a critical area would be required for runoff generation or saturated area (O'Loughlin, 1981, 1986).

The values of α_s and β_s, coefficient of determination (R^2) and significant level (p) were calculated for each observation of total nine cases (Table 9). The value of α_s (slope of regression line) means the ratio of discharge increment to drainage-area increment. If discharge increases in response to rainfall, α_s also increases. The values of α_s for storm-flow condition ($\alpha_s > 0.3$) are much larger than those for base-flow condition ($\alpha_s > 0.04$, Table 9).

Both of R^2 and β_s were plotted against the antecedent precipitation index (API_{30}) calculated by Equation (3), in order to analyze the variation of regression line in response to runoff conditions (Figure 28). The antecedent precipitation index (API_{30}) was used here as a parameter representing runoff condition. In Equation (5), the value of β_s is drainage area for $Q_s = 0$ (equivalent to the intercept by X-axis). In other words, β_s is equivalent to the critical area for runoff generation. The value of β_s decreases with increasing API_{30} (Figure 28). In the largest storm flow on 11 July, β_s declined to about 200 m^2 (Table 9 and Figure 27a).

Coefficient of determination (R^2) increases with increasing API_{30}, and R^2 rises to 0.79 in the storm flow on 11 July (Table 9). This result indicates that spring discharge is strongly controlled by drainage area in the storm-flow condition rather than in the base-flow condition.

As described in the section IV-4-2, the results of subsurface water response revealed that temporary groundwater table appeared at the bedrock-regolith boundary in storm runoff events (Figure 20). On the assumption of impermeable bedrock, the subsurface water, which cannot infiltrate into the bedrock, temporarily forms groundwater table at the bedrock-regolith boundary. Although the data on the permeability of bedrock with fractures are not available, high positive pressure heads immediately after rainfall peak indicate the generation of subsurface storm flow on the bedrock-regolith boundary (Figure 20).

The amount of subsurface storm flow at a point is generally controlled by the source area calculated from topography of bedrock-regolith boundary (Freer *et al.*, 1997, 2002). Hutchinson and Moore (2000) observed subsurface discharge from a trench at a slope with shallow regolith (1 m in depth), and suggested that the distribution of throughflow is affected by bedrock-regolith boundary for base-flow condition, but that throughflow is controlled by surface topography for storm-flow condition. Since regolith in the investigated area is thin (about 1 m), topography of bedrock-regolith boundary would approximate to surface topography. Moreover, in the case that all springs are supplied by subsurface storm flow, spring discharge should increase with increasing drainage area calculated from surface topography. This topographic control on subsurface storm flow should yield a better area-discharge correlation in springs for the storm-flow condition (Figure 28).

Temporary groundwater table fell down after rain stopped, and disappeared in a day. Thus, spring positions should also vary in response to the emergence of temporary groundwater table. The result from the temporal variation of spring positions indicates that the emergence and disappearance of subsurface storm flow in regolith would affect the spring positions (Figure 18).

The coefficients of determination for area-discharge relation were relatively low (0.3 – 0.4) in base-flow condition. Onda (1994a) and Komatsu and Onda (1996) found no correlation between base-flow discharge and drainage area in basins underlain by sedimentary rocks and serpentinite. In the case that springs are strongly affected by bedrock groundwater, stream discharge would not depend on drainage area, because the bedrock groundwater flow is controlled by internal structure such as the fractures, which is independent of surface topography. For example, Genereux *et al*. (1993) suggested that the spatial distribution of stream flow in a basin underlain by dolomite (West Fork of Walker Branch Watershed) is strongly affected by bedrock structure rather than surface topography. In the investigated area, spring discharge in base-flow condition would be controlled by bedrock groundwater flow through the fractures, which are not affected by surface topography.

Considering the runoff conditions for channel initiation, the storm-flow condition is more appropriate than the base-flow condition. As discussed later, the sediment transport primarily occurs under storm-flow condition; hence, a linear area-discharge relationship under storm-flow condition must be applied to runoff model in the investigated area.

V-2: Relationship between rainfall and peak discharge

Peak discharge in storm runoff must be the most important hydrological factor for sediment transport and channel initiation. The peak discharge is expressed by a function of drainage area and rainfall intensity in runoff prediction models for large-scale basins (Rodriguez-Iturbe, 1993). However, few studies investigated peak discharge at channel heads on the basis of physical theories. This section explores to set up the rainfall-runoff equation showing relationship among rainfall intensity, drainage area and peak discharge with the rainfall-runoff data at channel heads.

Expanding Equation (5), it is assumed that peak discharge, Q_p, in storm-runoff condition is linearly proportional to drainage area, A. This assumption yields:

$$Q_p = \alpha_p (A - \beta_p) \tag{6}$$

where α_p is the ratio of peak-discharge increment to drainage-area increment, and β_p is the critical drainage area for peak-runoff generation.

As described in the section V-1, the value of β_s decreases with an increase of antecedent rainfall (Table 9). The value of β_s decreased to 200 m^2 for the largest storm flow (Figure 27a). Since β_s at the recession stage of the largest storm flow was only 200 m^2, the value of β_p at the peak of the storm flow would be less than 200 m^2. In this case, the critical area would be enough smaller than the source area ranging from 530 to 16,900 m^2 (Table 2). In the following discussion, β_p is assumed to be negligible ($\beta_P \approx 0$). This assumption, therefore, yields:

$$Q_p = \alpha_p A \tag{7}$$

The value of α_p increases with increasing magnitude of rainfall. In the case that α_p is linearly proportional to rainfall intensity, I_R, i.e., $\alpha_p = k_p I_R$, Equation (7) can be rewritten by:

$$Q_{\mathrm{p}} = k_{\mathrm{p}} I_{\mathrm{R}} A \qquad (8)$$

where k_{p} is a dimensionless coefficient on runoff peak generation. Rainfall intensity, I_{R}, has the dimension of velocity. In the following discussion, since the rainfall intensity is considered to be effective for runoff-peak generation, I_{R} is denoted as effective rainfall intensity. Various time spans of rainfall are available for the calculation. Effective rainfall intensity, I_{R} (m s^{-1}), for T-hour rainfall can be expressed by:

$$I_{\mathrm{R}} = \frac{R_T - R_{\mathrm{c}T}}{3600T} \qquad (9)$$

where R_T (m) is the maximum rainfall amount for T hours in an event, and $R_{\mathrm{c}T}$ (m) is a critical T-hour rainfall for runoff generation. The unit of I_{R} in Equation (9) is m s^{-1}. Interception by forest canopy or soil moisture deficit prevents the increase of discharge for minor rainfall events (Beven and Kirkby, 1979); hence, a critical rainfall, $R_{\mathrm{c}T}$, was taken into account in Equation (9). Substitution of Equation (9) to Equation (8) yields the following rainfall-runoff equation:

$$\frac{Q_{\mathrm{p}}}{A} = \frac{k_{\mathrm{p}}}{3600T}\left(R_T - R_{\mathrm{c}T}\right) \qquad (10)$$

Least squares linear regression between peak specific discharge, Q_{p}/A, and T-hour rainfall, R_T, yields the suitable coefficients in Equation (10). All rainfall-runoff events at channel-head sites (C1L and C3U) for three years (2000 – 2002) were used for the regression analysis. A rainfall-runoff event was defined as an increase in discharge in response to the continuous rainfall, which can include no-rain periods within six hours. Since discharge rarely increased when rainfall is less than 10 mm, rainfall events with total rainfall of < 10 mm were not used for the analysis. If a no-rain period exceeds 6 h, two continuous rainfalls before and after the no-rain period were regarded as two distinct rainfall events. This criterion of 6-hour no-rain period is based on the following reasons. In the case of a shorter no-rain period (e.g. 1 h) than 6 h, rainfall of the 'previous' event may frequently affect the next runoff event. In the case of a larger no-rain period (e.g. 24 h) than 6 h, two or more isolated runoff peaks may be counted as the 'same' event.

Table 10 shows the dataset of T-hour rainfall and peak discharge used for the analysis. The maximum rainfall, R_T, was calculated for various durations, T, ranging from 0.167 h (10 min) to 48 h. In the case of $T > 6$ h, the maximum T-hour rainfall of an event occasionally includes part of the rainfalls during the previous event.

Figure 29 shows relationship between coefficient of determination, R^2, and duration of maximum rainfall, T. Coefficient of determination decreases below 0.8 when T is less than 2 h or more than 24 h. Peak specific discharge was plotted against maximum 1-hour rainfall (Figure 30) and maximum 24-hour rainfall (Figure 31). Although both Figures 30 and 31 reveal that peak specific discharge increases with increasing rainfall on the whole, these plots show a significant scattered distribution. The data at C1L and C3U show a similar distribution.

By contrast, R^2 exceeds 0.8 in the case between $T = 3$ h and 12 h in both C1L and C3U sites. The duration, T, which maximizes R^2, is 8 h for C1L site, and 4 h for C3U site. However, the difference in R^2 between C1L and C3U sites is small enough for $3 < T < 12$ h. The combined data for C1L and C3U sites (squares in Figure 29) show that the R^2 is maximized when $T = 4$ h. Plots of peak specific discharge against maximum 4-hour rainfall show clear correlation (Figure 32). Therefore, the following regression line calculated from maximum 4-hour rainfall, R_4, represents the rainfall-runoff relation at channel heads in the investigated area:

$$\frac{Q_p}{A} = 68.7 \times 10^{-6} (R_4 - 0.014) \qquad (R^2 = 0.84) \qquad (11)$$

Comparing Equations (10) and (11), the following values were obtained: $k_p/3600T = 68.7 \times 10^{-6}$ s^{-1} and $R_{c4} = 0.014$ m.

V-3: Critical discharge for bedload transport

Bedload yield and peak discharge in each period at C1L, C3U and C3L sites were summarized in a dataset (Table 11). In order to estimate critical discharge for bedload transport in channel heads, the bedload yields at C1L and C3U sites were plotted against the peak discharges (Figure 33). Figure 33 do not include data in the winter when hydrological observation was suspended (mainly from November to May, Table 7). Bedload yields in the periods without storm-runoff events (noted as NE in Table 11) are also excluded. Figure 33 indicates that when the peak discharge exceeded a critical value, bedload yield abruptly increased at both C1L and C3U sites. The critical discharges for bedload transport, Q_{cr}, were estimated from the two dashed lines in Figure 33 to be 0.035 m^3 s^{-1} at C1L site and 0.007 m^3 s^{-1} at C3U site.

In general, critical discharge, Q_{cr}, increases with decreasing channel gradient, S_c.

The relationship between Q_{cr} and S_c can be expressed by a power function:

$$Q_{cr} = \gamma\, S_c^{-m} \qquad (m > 0) \qquad (12)$$

where γ and m are the constants depending on flow condition. Montgomery and Dietrich (1994b) suggested the equations of $Q_{cr} - S_c$ relation for both laminar-flow and turbulent-flow conditions. Critical discharges in the turbulent-flow condition, Q_{crT}, and in the laminar-flow condition, Q_{crL}, are expressed by:

$$Q_{crT} = \frac{\gamma_T}{S_c^{7/6}} \qquad (13)$$

$$Q_{crL} = \frac{\gamma_L}{S_c^{2}} \qquad (14)$$

$$\gamma_T = \frac{b\tau_{cr}^{5/3}}{n(\rho_w g)^{5/3}} \qquad (15)$$

$$\gamma_L = \frac{2b\tau_{cr}^{3}}{k\nu\rho_w^{3} g^{2}} \qquad (16)$$

where b is channel width, τ_{cr} is critical shear stress for bedload transport, ρ_w is density of water, g is gravitational acceleration, n is Manning's resistance coefficient, k is surface roughness coefficient, ν is kinematic viscosity, and γ_T and γ_L are constant on critical discharge in turbulent-flow and laminar-flow conditions, respectively.

Figure 34 shows relationship between critical discharges, Q_{cr}, against channel gradient, S_c, for the observation sites. Channel gradient is local gradient from an observation site to the point 10 m upstream. The values of S_c at C1L and C3U sites are 0.384 (21.0°) and 0.758 (37.2°), respectively (Table 4).

Empirical $Q_{cr} - S_c$ relation can be estimated from substitution of Q_{cr} and S_c values at both C1L and C3U sites into Equation (12). The result yields $\gamma = 0.0036$ $m^3 s^{-1}$ and $m = 2.37$, i.e.:

$$Q_{cr} = 0.0036 S_c^{-2.37} \tag{17}$$

Flow condition immediately below the channel head is unknown. Substitution of Q_{cr} and S_c at each site into Equations (13) or (14) yields the thresholds for bedload transport on the assumption of turbulent-flow or laminar-flow conditions. The dashed lines and thin solid lines in Figure 34 indicate the thresholds for bedload transport in turbulent-flow and laminar-flow conditions, respectively. Thresholds for bedload transport in turbulent-flow condition at two sites are apart each other (two dashed lines in Figure 34). The value of γ_T at C1L (0.0115 $m^3 s^{-1}$), which is calculated with Equation (13), is about twice that at C3U (0.0051 $m^3 s^{-1}$). Although two γ_L values in laminar-flow condition give the better agreement (two thin solid lines in Figure 34), there is a little difference between the γ_L at C1L (0.0052 $m^3 s^{-1}$) and C3U (0.0040 $m^3 s^{-1}$). Thus, Equation (17) is the most suitable $Q_{cr} - S_c$ function in the investigated area.

V-4: Thresholds for channel initiation by bedload transport

V-4-1: Equations

This section proposes a new method to estimate the thresholds for channel initiation by bedload transport with hydro-geomorphic data. Discussion in this section is based on the following three assumptions: (1) peak discharge in a storm event is linearly proportional to drainage area, (2) peak discharge in a storm event is linearly proportional to maximum T-hour rainfall of the storm, and (3) a critical discharge for bedload transport exists. Following discussion is only applicable to the mountains, which satisfy the above assumptions. As discussed above, the hydro-geomorphic conditions in the investigated area satisfy these assumptions.

If peak discharge, Q_p, in a storm runoff increased to the critical discharge for bedload transport, Q_{cr}, at a site, i.e., when $Q_p = Q_{cr}$, bedload transport occurs. Thus, thresholds for bedload transport can be rewritten through combining the rainfall-runoff equation (Equations 8 or 10) with critical discharge (Equation 12):

$$AS_c^{\,m} = \frac{\gamma}{k_p \cdot I_R} = \frac{\gamma \cdot 3600T}{k_p \cdot (R_T - R_{cT})} \tag{18}$$

This equation indicates general thresholds for bedload transport under a given gradient, source area, and rainfall condition. Discussion in the section V-2 revealed that maximum 4-hour rainfall R_4 (T = 4 h) is available for effective rainfall intensity. Substitution of $k_p/3600T = 68.7 \times 10^{-6}\ \mathrm{s^{-1}}$, $R_{c4} = 0.014$ m, $\gamma = 0.0036\ \mathrm{m^3\ s^{-1}}$, and $m = 2.37$ into Equation (18) yields:

$$AS_c^{\,2.37} = \frac{52.4}{R_4 - 0.014} \tag{19}$$

Equation (19) shows the thresholds for bedload transport in the investigated area. The rainfall condition required for the bedload transport at channel heads can also be calculated from topographic data of each channel head. Transforming

Equation (19) yields:

$$(R_4)_{cr} = \frac{52.4}{AS_c^{2.37}} + 0.014 \tag{20}$$

Substitution of the $AS_c^{2.37}$ values calculated from topographic data of each channel head into Equation (20) yields the $(R_4)_{cr}$ values for the channel head. Statistical distribution of the $(R_4)_{cr}$ in an area may be affected by the distribution of grain size or surface roughness in the area (Istanbulluoglu *et al.*, 2002). In the present study, the distribution of k_p and R_{c4} also contributes to the statistical distribution of the $(R_4)_{cr}$. However, on the assumption that the values of k_p, R_{c4}, γ and m are spatially uniform for all channel heads in the area, the R_4 value calculated in Equation (20) means the 4-hour rainfall required for bedload transport at channel heads. In this sense, the value of R_4 calculated by Equation (20) should be referred to as 'critical 4-hour rainfall, $(R_4)_{cr}$'.

V-4-2: Channel heads and calculated thresholds

Figure 35 shows relationship between channel gradient and source area in 24 channel heads. The three lines in Figure 35 indicate the thresholds for bedload transport in the three rainfall cases: R_4 = 40 mm, 90 mm, and 200 mm. Considering the condition of bedload transport at the points immediately below the channel heads, channel gradient instead of head slope is suitable for the horizontal axis. Figure 35 shows that most of the channel heads are plotted around the threshold line of R_4 = 90 mm. Estimation of return period suggests that the storm of R_4 = 90 mm, which corresponds to the rainfall condition for bedload transport in most channel heads, will occur once in three years (see Appendix B).

Figure 36 shows the statistical distribution of $(R_4)_{cr}$ values in 24 channel heads. The average of $(R_4)_{cr}$ is 95.8 mm and standard deviation, σ, is 52.7 mm. The minimum of $(R_4)_{cr}$ is 33 mm, and the number of channel heads with $(R_4)_{cr}$ of < 50 mm is six (25 % of all). In these six channel heads, bedload transport should occur several times in a year (see Appendix B). As described below, however, the condition of sediment in three of these channel heads (O27H, O51H, and O55H) is particular; hence, the values of k_p and γ are different with those of the investigated watersheds (C1 and C3). For example, the channel head, O27H, have the sediment made of quite larger grain size (about 1 m in diameter) than the investigated sites (C1L and C3U). The channels in O51 and O55 basins have no bedrock exposure on the whole channel section, and the channel-bed sediment is thicker than that in C1 and C3 watersheds (see Figures 13 and 17). Thick sediment accumulation may reduce the discharge due to groundwater flow beneath the channels. Four-hour rainfall for bedload transport in three channel heads (O27H, O51H, and O55H) may be underestimated because of the variety of channel-head condition.

In contrast, bedload transport in seven channel heads (28% of all) requires the critical 4-hour rainfall $(R_4)_{cr}$ of > 120 mm (Figure 36). The maximum value of $(R_4)_{cr}$ is 249 mm. Although the return period of R_4 = 120 mm is about 10 years, the return period of the maximum rainfall (R_4 = 249 mm) exceeds 200 years (see Appendix B). Bedload transport in four channel heads (17 % of all) requires infrequent storms with a return period of over 30 years (R_4 > 143 mm, Figure 36). However, rockfalls occur frequently in the intervals of the bedload transporting events. The rock fragments supplied by rockfall would accumulate on the channel heads, burying the topography of the channel heads. If the interval becomes longer, the role of the rock accumulation would increase.

Threshold for shallow landsliding cannot be estimated in the investigated area, since the high content of gravels (50 – 80%) prevents the measurement of the shear strength of regolith. Consequently, the present study cannot evaluate the role of shallow landslides, which are considered to be important for channel initiation and development of zero-order basins (Tsukamoto *et al*., 1982; Dietrich *et al*., 1986). Although shallow landslides appear to be infrequent in the investigated area underlain by Mesozoic bedded chert, small landslide scars are found in the five channel heads (Table 2). For example, the channel heads in C3 watershed (C3H, Figure 13 and Photo 5) has a small landslide scar with 5 m in width. A field observation suggests that shallow landslides also affect the channel initiation in some steep (Type-S) channel heads. This result concurs with the results by Dietrich *et al*. (1992, 1993), who suggested that the steeper channel heads are more sensitive to the shallow landslides than the gentler channel heads.

Relatively frequent rainfall with the return period of less than 30 years is required for bedload transport in 20 channel heads (83 % of all). Most of the channel heads in the investigated area are located where bedload transport occurs in the relatively frequent (< 30-year) rainfall. Thresholds for bedload transport rather than shallow landsliding should control the threshold for channel initiation in most of channel heads.

Survey of channel-head location as well as observation in the investigated watersheds provides the evidence of frequent bedload transport. As described in the section IV-5-2, distinct evidence of bedload transport was obtained in C3 watershed (Figure 22). Although the total bedload yield was smaller in C1 watershed, a distinct event of bedload transport occurs at least once at C1L site over three years (Figure 33). Field survey of channel-head locations provides the qualitative evidence. On the first-order channel immediately below O19H channel heads, a recent removal of debris was found on 30 November in 2000. A recent deposition of debris was also found at the mouth of O52 basin on 22 September in 2001. This new deposition would be simultaneously generated in the Period I (5 August to 1 September) of 2001 when the largest bedload yield was observed at C3U and C3L. These observations support the frequent occurrence of bedload transport on first-order channels immediately below the channel heads.

V-5: Effect of type of channel head on erosion rate

Coupling the type of channel head with total bedload yield in first-order basins, quantitative discussion will be possible concerning the effect of channel head on erosion rate. Although precise erosion rate can be calculated from long-term observation of bedload, suspended, and solute yield (Hirose, 1996), total bedload yield for three years (2000 – 2002) is assumed to represent erosion rate in the following discussion.

Total bedload yield at C3U (> 298 kg) was at least 20 times larger than that at C1L (14 kg, Table 8). Figure 37 shows relationship between local channel gradient, S_c, and annual bedload yield per unit area, Y_t/A. Annual bedload yield per unit area at C3U (> 0.0585 kg m^{-2} y^{-1}) was at least 80 times larger than that at C1L (0.0007 kg m^{-2} y^{-1}, Table 8). Both basin relief ratio and local channel gradient at C3U are larger than those at C1L, and consequently, erosional processes at C3U seem to be more active than that at C1L.

Although local channel gradient at C3L is almost same as that at C1L, annual bedload yield per unit area, Y_t/A, at C3L (> 0.0423 kg m^{-2} y^{-1}) was at least 60 times larger than that at C1L (0.0007 kg m^{-2} y^{-1}, Figure 37). Both C3U and C3L sites had larger bedload yield per unit area than C1L sites. This result implies that the whole channel sections of C3 watershed have a larger erosion rate than C1 watershed. Bedload transport on first-order channels may be controlled by bedload transport at the channel heads rather than local channel gradient of each site.

As described in the section III-2, two types of channel heads (Types G and S) exist in the investigated area. The channel head in C1 is a Type G with a larger source and a gentler channel gradient, whereas the channel head in C3 is a Type S with a smaller source and a steeper channel gradient. Thus, the result of bedload yield at C1L and C3U sites imply the faster landform evolution in Type-S channel heads than Type-G channel heads. Discussions in this section, however, are based on short-term data (only three years), and observation sites are spatially limited. The other methodologies to estimate erosion rate as well as the long-term bedload measurement at widely distributed sites will be required for quantitative analysis for landform evolution of channel heads.