The role of soil creep and slope failure in the landscape evolution of a head water basin: field measurements in a zero order basin of northern Japan

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Abstract

Field measurements of soil creep and slope stability were conducted on a nose, side-slope and hollow in a zero order basin near Sapporo, Hokkaido, northern Japan, and the preferential location of soil creep and slope failure was determined. Soil creep was continuously measured by the strain probe method at three sites from 1994 to 1995, and was compared with soil moisture conditions and ground temperature. In the summer, active soil creep occurred only when rainfall led to large soil moisture changes and a near-saturated condition, which was most likely induced by shrink–swell activity of soil. In the winter, soil creep was caused by seasonal frost, although the mass transport was limited because of the insulation provided by snow cover. These results indicate that the soil moisture change and soil moisture content during a rainfall event in the summer are the major factors controlling soil creep in this basin. Soil moisture conditions were further measured by a tensiometer at 16 sites in the rainy season in 1994. On the nose and side-slope, active soil-moisture changes took place during rainfall-events. The hollow tended to maintain higher soil-moisture conditions than the nose and side-slope, because subsurface flow was concentrated in the hollow. Thus the soil-moisture variation that encourages soil creep rarely occurred in the hollow. From these results, sediment transport rates caused by creep were estimated to be $2.07 \times 10^{-3}$ m$^3$/yr on the nose, $1.595 \times 10^{-3}$ m$^3$/yr on the side-slope and $9.0 \times 10^{-3}$ m$^3$/yr on the hollow, respectively. These results clearly show infilling in the hollow and denudation on the nose. Slope stability was analyzed by the infinite slope model. The potential of slope failure was evaluated from the relationship between critical water depth $H_c$ and soil thickness $D$. The analysis revealed that an increase in $D$ causes a marked decrease in $H_c$ on the side-slope, indicating the high potential of slope failure on the slope. In contrast, both on the nose and in the hollow, the decrease in $H_c$ for the same increase in $D$ was lower than that on the side-slope. However, slope failure on the side-slope and soil creep on the nose infill material into the hollow. Thus, the increase in $D$ in the hollow is higher than that on the other slopes; leading to an increase in slope failure potential. These results indicate that soil creep and slope failure act as infilling and evacuating processes of the zero order basin with differing intensities depending on slope form: soil creep removes soil materials from the nose and deposits them in the hollow, whereas slope failure removes materials from the

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side-slope and deposits them in the hollow. When infilling develops a sufficiently thick soil accumulation in the hollow, slope failure evacuates the hollow. © 1999 Elsevier Science B.V. All rights reserved.

**Keywords:** zero order basin; hollow; colluvium; soil creep; slope failure

### 1. Introduction

This study examines soil creep processes and slope stability, and focuses on the preferential location of soil creep and slope failure in a zero order basin. A zero order basin is usually composed of three slope forms, i.e., nose, side-slope and hollow (Hack and Goodlett, 1960). Many geomorphologists have paid particular attention to the hollow because thick colluvium, which is the primary source of slope failures and resultant debris flows, is infilled into the hollow (Dietrich and Dunne, 1978; Dietrich and Dorn, 1984; Marron, 1985; Dietrich et al., 1986; Reneau et al., 1986, 1989; Reneau and Dietrich, 1987, 1990, 1991; Crozier et al., 1990; Fernandes et al., 1994; Yamada, 1995; Yoshinaga and Koiwa, 1996). As a result of the numerous studies mentioned above, the distribution, depositional age and depositional rate of colluvium are becoming increasingly well model. However, the processes supplying the colluvium to the hollow are poorly understood. Since the source areas are nose and side-slope, processes operating on these slopes are important in the landscape evolution of headwater basins. Thus, field measurements on soil-creep processes and slope stability were conducted on a nose, side-slope and hollow in a zero order basin near Sapporo, Hokkaido, northern Japan.

### 2. Study area

#### 2.1. Physiography

The study area is located in a forested, north-facing zero order basin in the Hiyamizusawa Brook basin, 20 km southwest of Sapporo, Hokkaido, northern Japan (Fig. 1). This zero order basin has an

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Fig. 1. Study area. Triangles are the measurement sites for pressure head, and triangles with R, S and V are the measurement sites for soil creep. The solid circle with RG is the measurement site for precipitation.
area of 0.02 km² and ranges from 500 to 800 m a.s.l., and is underlain by Tertiary andesite Doi, 1953. The slope is covered with a forest of broadleaf trees such as Acer mono Maxim. and Viburnum furcatum Blume, and coniferous trees such as Picea jezoensis Carr., and the forest floor is covered with undergrowth such as Maianthemum dilatatum. Annual precipitation is ca. 1400 mm. The total precipitation during the relatively dry period from May to mid-August is 100 to 150 mm, in contrast to that of the late summer rainy season, which ranges from 400 to 500 mm. Evidence of overland flow has not been observed on the slopes in this basin even in the rainy season. The period of November to March is the snowy season, and snow mantles the basin from December to April. The mean air temperature of Sapporo is 8°C.

2.2. Slope form and soil distribution

This zero order basin can be subdivided into nose, side-slope and hollow (Fig. 1). The nose includes the ridge crests and nearby areas in which the contours are convex-outward. The side-slope is located on the inner side of the nose, and the contours are straight or nearly so. The hollow occupies the central part of the basin, and the contours are concave outward.

The soil distribution in the zero order basin clearly shows that the colluvial soils are infilled into the hollow, and the materials infilling the hollow were supplied from the nose and side-slope (Fig. 2). The colluvial soil infilling the hollow ranges thickness from 60 to 180 cm. A and C horizons are 10 to 30 cm and 40 to 150 cm thick, respectively, whereas the B horizon is scarcely developed. Buried horizons were observed below the C horizon, although the thicknesses are only 10 to 30 cm. Gravel content is 30% in both A and C horizons, and sand comprises 70 to 90% of the fine material. At some locations in the hollow, gravelly colluvial soil without fine materials was found. In contrast to the colluvial soils in the hollow, very thin soil (regosol), thinner than 40 cm, is distributed on part of the nose and a large part of the side-slope. The regosol has developed only A and C horizons and includes 20 to 40% gravel. Denudation of soil materials results in these immature regosol features. Thus, the distribution of regosol shows denudation on a large part of the side-slope and part of the nose. The materials denuded on those slopes would be infilled as the colluvial soil in the hollow, which is located in the lowermost part of the basin. The rest of the basin, that is, the greater part of the nose and part of the side-slope, is covered with residual soil 50 to 130 cm thick. The residual soil has A, B and C horizons and contains 0 to 10% gravel. The fines in this soil consist of 85 to 90% sand.

3. Soil creep processes

3.1. Methodology

Soil creep was continuously measured by the strain probe method at three sites, and was compared...
with soil moisture conditions and ground temperature (Fig. 1). The strain probe is constructed of a spring steel strip 1.3 cm wide, 0.4 mm thick and 50 or 40 cm long. Two pairs of strain gauges of 3 cm long were attached to the strip in a 10 cm section, and were combined to construct a wheatstone bridge circuit (Fig. 3). The strip was coated with silicone rubber to protect the strain gauges. The strain probe was initially inserted perpendicular to the ground surface. Since the soil creep varies with depth from the ground surface (e.g., Carson and Kirkby, 1972), the inserted probe will be bent. The amount of bending at various points on the probe was evaluated from a change in the measured strain of the gauges. Soil displacement was calculated from the strains, on the assumptions that each section of the probe bends circularly, and the circular arc of each section connects continuously to that of the next section. The formulae for calculating the displacement are described in detail in Yamada and Kurashige (1996).

The strain value was recorded automatically or manually. At site R, strain was recorded automatically in a data logger at hourly intervals. At sites S and V, the strain was measured manually with a strain meter at 5-day intervals and also on rainy days. The length of the probe was 50 cm at sites R and V, and 40 cm at site S.

The pressure head in the soil, ground surface temperature and precipitation were also measured. Pressure-heads at 30 cm depths were measured on the nose, side-slope and hollow by a tensiometer at 16 sites including sites R, S and V (Fig. 1). The measurements were conducted at 5-day intervals, including rainy days, from July to October 1994, which is the rainy season in this area. Ground surface temperature was measured with a thermistor at site R at 1 h intervals. Precipitation was measured at 15-min intervals by a tipping-bucket rain gauge at site RG (Fig. 1). These measurements were carried out from July to October 1994, and the strain and

![Fig. 3. Arrangement of the strain probe and bridge circuit. Two pairs of strain gauges 30 mm long were attached to the strip in a 100-mm section, by which all four strain gauges cover 60% of the length of the 100-mm section.](image)
ground temperature at site R were measured until May 1995.

3.2. Factors of soil creep

In the summer and autumn, obvious soil displacement occurred during the rainfall events in the period before mid-August at all sites (Fig. 4). On August 3, the soil was displaced downward, and subsequently retrograded. The retrograde displacement continued for a week, and cancelled the downslope displacement. From August 13 to 15, downslope displacement occurred but the subsequent retrogression was negligible. The downslope displacement during the event was approximately 1 mm at the ground surface at all sites, decreasing downward, and being negligible deeper than 40 cm (Fig. 5). After mid-August, although intense rainfall occurred, significant soil displacement did not take place at all sites.

The rainfall caused significant change in pressure head in the period before mid-August, when soil displacement occurred. Since the rainfall in mid- and late July was negligible (1.0 mm), the pressure head was lowest in late July. The lowest value was $-555$, $-723$ and $-195$ cm H$_2$O at sites R, S and V, respectively. The rainfall, caused significant change in pressure head in the period before mid-August, when soil displacement occurred. Since the rainfall in mid- and late July was negligible (1.0 mm), the pressure head was lowest in late July. The lowest value was $-555$, $-723$ and $-195$ cm H$_2$O at sites R, S and V, respectively.
respectively. During the rainfall event on August 3 (11.5 mm), the pressure head increased to $-212$, $-673$ and $-71$ cm H$_2$O at sites R, S and V, respectively, and subsequently decreased to the same level as that of late July. A greater change in the pressure head occurred during the rainfall event from August 12 to 14 (109.5 mm). The value rose from the lowest value to a near-saturated level ($>-50$ cm H$_2$O) at all sites, and subsequently fell to $-217$, $-574$ and $-74$ cm H$_2$O at sites R, S and V, respectively.

In contrast, after the rainfall in mid-August, although intense rainfall occurred frequently, the change in the pressure head during a rainfall event was small. Intensive rainfall events in amounts exceeding 20 mm per day, took place six times in September, and the pressure head rarely fell below $-140$ cm H$_2$O at all sites after late September. The frequent rainfall allowed the soil to remain wet, and prevented significant changes in the pressure head.

These data suggest that the significant soil-moisture change caused the soil displacement, as is clearly shown in Fig. 6. The large soil displacement ($D_G \geq 0.5$ mm in Fig. 6) occurred only when the change in pressure head during the rainfall event was relatively large. Even if the soil was near saturation, the soil displacement was small when the pressure head change was small. It is well known that shrink–swell activity of soil, induced by soil moisture change, generates downslope soil displacement, because overburden pressure displaces the soil downhill during a shrink–swell cycle (Young, 1960; Kirkby, 1967). Shrink–swell is intensified if the soil contains swelling clay minerals, such as smectite (Selby, 1993). In this basin, the soil of the hillslope includes swelling smectite (Kurashige, 1993). Thus, the downslope displacement is considered to have been generated by shrink–swell of soil, which was in turn induced by the large soil moisture change.

Fig. 6 also shows that retrograde displacement occurred when soil-moisture change was relatively large, but the pressure head during the rainfall event was well below the near saturated condition. Retrogression can appear as a result of soil displacement perpendicular to the ground surface if the probes were not inserted completely normal to the ground surface. Actually the probes were probably inserted into the soil inclined at a small angle to the normal of the surface ($\beta$ in Fig. 7), because it is difficult to insert them completely normal to the surface. When the angles of shrinking and swelling ($\gamma_{sw}$, $\gamma_{sh}$ in Fig. 7) are less than $\beta$, the downslope displacement and the upslope retrogression can be detected during the swelling and shrinking of soil, respectively (Fig. 7).

![Fig. 6. Types of soil displacement relative to soil moisture conditions. The change in pressure head during a rainfall event, $\phi_c$, and the pressure head during a rainfall event, $\phi_r$, are shown in the schematic diagram.](image-url)
Retrograde displacement, however, becomes small when $\gamma_w$ and $\gamma_h$ are large. The retrogression was large when the soil moisture was well below the near-saturated condition during a rainfall event, and was small when the soil was nearly saturated during an event (Fig. 6). Thus, angles $\gamma_w$ and $\gamma_h$ for the near-saturated condition were judged to be larger than that for the low soil-moisture condition. Since vadose water with low pressure head, i.e., high soil water tension, adheres soil grains strongly to one another, the soil elasticity generated by the tension might result in small $\gamma_w$ and $\gamma_h$.

Net downslope displacement during a shrink–swell cycle becomes large when both the amount of shrink–swell and $\gamma_w$ and $\gamma_h$ are large. The amount of shrink–swell increases with increasing soil moisture change during a rainfall event, and $\gamma_w$ and $\gamma_h$ become large when the soil is nearly saturated during a rainfall event. Consequently, it is concluded that soil moisture content and its change during a rainfall event are important for soil creep in summer.

Downslope displacement in winter and spring was also examined to consider whether soil creep was induced by the freeze–thaw activity of the soil. Fig. 8 shows the displacement and ground surface temperature at site R from November 1994 to May 1995. The precipitation data, obtained at Koganeyu Station located 6 km northeast of the study area (Sapporo District Meteorological Observatory, 1994–1995), are also illustrated. The total precipitation at Koganeyu Station was 650 mm during the entire period, and the precipitation from December 1994 to March 1995 was 361 mm, most of which was snow. Snow depth was not monitored, but snow of 30 cm depth was measured at site R on 19 April 1995.

In winter and spring, gradual downslope displacement and abrupt retrograde displacement occurred with the seasonal freeze–thaw cycle. The gradual downslope displacement began in mid-December with the onset of the seasonal frost, and continued until early April. Although the downslope displacement at the ground surface reached 1.2 mm during the seasonal frozen period, it was followed by an abrupt retrograde displacement in late April concurrent with the spring thaw.

The depth of soil displacement, which was only the upper 10 cm, coincides with the frost penetration depth. The frost penetration depth $D$ is given by

$$D = a\sqrt{F}$$

where $a$ is a constant that usually ranges between 2 and 3, and $F$ is the freezing index (Kinoshita, 1984). The value of $F$ was estimated to be 26.2°C days from the ground surface temperature data. Setting $a = 2$ leads to $D = 10$ cm, which agrees with the depth of soil that is gradually displaced downward.

It is well known that the freezing of soil displaces the soil perpendicular to the ground surface; this process is called frost heave. This type of soil displacement deforms the probe if the probe is not inserted completely normal to the ground surface (see Fig. 7). Thus, it is likely that the seasonal frost heave deformed the probe section from the upper
The precipitation was recorded at Koganeyu Station located 6 km northeast of the study area.

10 cm to downslope. This idea is supported by the correlation of the soil displacement depth and the frost penetration depth. A heaved soil layer is generally displaced downslope with thawing of the soil (e.g., Higashi and Corte, 1971). However, the obtained data showed that almost all the downslope displacement measured by the strain probe during the seasonal freezing was cancelled by the spring thaw. A certain force is needed to deform the probe because it is an elastic body. If the heaved layer is thin, the driving force is weak. The heaved layer seems to be limited to the surface layer, because of the shallow frost penetration. This suggests that the driving force during the spring thaw was weak. Consequently, it is likely that downslope soil displacement actually occurred during the spring thaw but the force exerted by the thin heaved layer was insufficient to allow the probe to record the downslope deformation.

The snow cover limits the soil displacement to a thin surface layer. At an experimental plot without snow cover in Sapporo, where the air temperature in winter is similar to that in the study basin, the depth of displacement is 42 cm (Fukuda and Ishizaki, 1980). In contrast, in this snow-covered basin, the frost depth was estimated to be only 10 cm because of the insulation provided by snow cover. The thin frost depth confined the frost heave to a thin layer. Thus, it is likely that the effect of snow cover on the frost depth resulted in only minimal mass-transport, and in turn prevented the probe from downslope deformation during the spring thaw.

3.3. Soil-moisture conditions on the nose, side-slope and hollow

The soil-moisture conditions in summer are the most important factor controlling soil creep in this basin. Although freeze–thaw activity induces frost creep, the resulting mass transport is very small. Thus, we can evaluate the preferential location of soil creep in terms of soil moisture conditions.
Pressure head tends to be lowest on the nose and highest in the hollow throughout the measurement period (Fig. 9). Pressure-heads significantly decreased from July to mid-August (approximately \(-500\) cm H\(_2\)O for nose and side-slope, \(-300\) cm H\(_2\)O for hollow), although a temporal increase induced by rainfall of 11.5 mm occurred on August 3. A large increase in pressure head on all slopes took place in the rainfall event from August 12 to 14 (total 109.5 mm), and after the event, the pressure head decreased to only \(-300\), \(-200\) and \(-70\) cm H\(_2\)O for the nose, side-slope and hollow, respectively. Then, the rainfall event on September 7 (total 21.5 mm) caused pressure-heads to increase on all slopes, and high pressure-heads (more than \(-100\) cm H\(_2\)O) were maintained for the period from mid-September to late October.

The change in pressure head during the major rainfall events on August 3, August 14 and September 7, was largest on the nose and smallest in the hollow (Fig. 10). On August 3, although the pressure head changed during and after the rainfall, the head did not reach \(-50\) cm H\(_2\)O, which is the critical value for soil creep. In contrast, the pressure head exceeded \(-50\) cm H\(_2\)O at almost all the sites on August 14 and September 7. On August 14, the amount of the change was 600 to 1200 cm H\(_2\)O at almost all sites on the nose and the side-slope, whereas it was only 0 to 600 cm H\(_2\)O at almost all the sites in the hollow. On September 7, the pressure-head change exceeded 300 cm H\(_2\)O at several sites on the nose, whereas the head hardly changed on the side-slope and in the hollow. These results indicate that the large soil-moisture change necessary for soil creep occurs on the nose and on the side-slope.

The soil-moisture conditions on the nose, side-slope and hollow depend on their topographic features. It is well known that subsurface flow concentrates on a slope with concave-outward contours (e.g., Kirkby and Chorley, 1967; Anderson and Burt, 1978; Whipkey and Kirkby, 1978; Burt and Butcher, 1985). Therefore, the hollow tends to maintain higher soil-moisture conditions than the nose and the side-slope. Dry antecedent conditions are most likely to occur on the nose, because the convex curvature favors the drainage of subsurface water. Accordingly, the soil-moisture change during a rainfall event becomes greatest on the nose.

**3.4. Sediment transport rates**

The annual sediment transport rates by soil creep on each slope were estimated from the results of the soil creep and soil-moisture measurements described above. The estimation is based on the following assumptions: (1) sediment transport rate per unit width, \(S\), is proportional to the change in the pressure head during a rainfall event when the pressure head during the event exceeded \(-50\) cm H\(_2\)O, that is,

\[
S = K \sum \phi_c
\]
Fig. 10. Spatial distribution of soil moisture conditions during the rainfall events on August 3, August 14 and September 7. The change in pressure head during a rainfall event, $\phi_c$, and pressure head during the event, $\phi_f$, are shown in the schematic diagram.

where $K$ is a constant; (2) the constant $K$ takes the same value on each slope, and is given by

$$K = \frac{S_s}{\phi_{ca}}$$

where $S_s$ and $\phi_{ca}$ are the sediment transport and pressure head change, respectively, which were measured during the rainfall event on August 14 at the three sites by means of strain probe and tensiometer, respectively; and (3) sediment transport occurs normal to the perimeter of each slope; that is, sediment transport rate, $Q$, on each slope is product of $S$ and perimeter length, $l$.

$$Q = Sl.$$  

Although sediment transport rate by soil creep would be controlled by various factors, only topographic and hydrologic factors are examined in the estimation. These two factors depend on slope form, while other factors such as climate and geology do not change in this small zero order basin. Thus, the relative amount of sediment transport rate is comparable along each slope.

Table 1

<table>
<thead>
<tr>
<th>Site</th>
<th>$S_s$ ($\times 10^{-4}$ m$^3$/m)</th>
<th>$\phi_{ca}$ (cm H$_2$O)</th>
<th>$K$ ($\times 10^{-7}$ m$^3$/m cm H$_2$O)</th>
</tr>
</thead>
<tbody>
<tr>
<td>R</td>
<td>2.71</td>
<td>754</td>
<td>3.6</td>
</tr>
<tr>
<td>S</td>
<td>1.80</td>
<td>1098</td>
<td>1.6</td>
</tr>
<tr>
<td>V</td>
<td>1.79</td>
<td>224</td>
<td>8.0</td>
</tr>
</tbody>
</table>

$S_s =$ sediment transport per unit width in the rainfall event on August 14. 
$\phi_{ca} =$ change in the pressure head during the rainfall event. 
$K = S_s / \phi_{ca}$.
Table 2
Summation of change in pressure head during a rainfall event, $\Sigma \phi$, for each slope

<table>
<thead>
<tr>
<th></th>
<th>$\phi_i$ on August 14 (cm H$_2$O)</th>
<th>$\phi_i$ on September 7 (cm H$_2$O)</th>
<th>$\Sigma \phi_i$ (cm H$_2$O)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nose</td>
<td>729</td>
<td>227</td>
<td>956</td>
</tr>
<tr>
<td>Side-slope</td>
<td>642</td>
<td>136</td>
<td>778</td>
</tr>
<tr>
<td>Hollow</td>
<td>326</td>
<td>47</td>
<td>373</td>
</tr>
</tbody>
</table>

$K$ estimated from Eq. (3) ranged between 1.6 and $8.0 \times 10^{-3}$ m$^3$/m cm H$_2$O, as shown in Table 1. The $K$ values vary among the sites. Thus the maximum, minimum and average values of $K$ are written as $K_{\text{max}}$, $K_{\text{min}}$ and $K_{\text{avg}}$, respectively, and these three $K$ values were used for further analysis.

$\Sigma \phi_i$, averaged for the nose, side-slope and hollow, was 956, 778 and 373 cm H$_2$O, respectively (Table 2), showing the largest value on the nose and the smallest in the hollow. The summation of $\phi_i$ was calculated for the rainfall events on August 14 and September 7, 1994, because $\phi_i$ was negligible and/or the pressure head did not exceed $-50$ cm H$_2$O during the other rainfall events in 1994. $\Sigma \phi_i$ was calculated for each of the 16 sites in which the pressure head was measured, and the average for each slope was used for further analysis.

The resultant $S$ calculated from Eq. (2) was the smallest in the hollow because it had the smallest soil-moisture changes (Table 3). Values of $S$ ranging from 0.6 to $3.0 \times 10^{-3}$ m$^3$/yr in the hollow were only about half of those on the nose and the side-slope (1.2 to $7.6 \times 10^{-3}$ m$^3$/yr).

In addition to the small $S$, the short perimeter resulted in extremely small values of $Q$ in the hollow (Table 3). $Q$ values, which range from 58.0 to $376.3 \times 10^{-3}$ m$^3$/yr on the nose and the side-slope were approximately 20 times greater than $Q$ values in the hollow ($3.3$ to $16.4 \times 10^{-3}$ m$^3$/yr). These results demonstrate that the sediment discharge by soil creep from the hollow is much smaller than that supplied from the nose and the side-slope.

The mass balance calculated for $Q_{\text{avg}}$ shows evi- dent infilling in the hollow (Fig. 11). Soil creep supplied sediment at $181.9 \times 10^{-3}$ m$^3$/yr into the hollow, whereas sediment discharge from the hollow was only $9.0 \times 10^{-3}$ m$^3$/yr. The resultant balance is $172.9 \times 10^{-3}$ m$^3$/yr. This balance shows that 95% of the sediment supplied from the nose and the side-slope was stored in the hollow.

The balance also shows the significance of denudation caused by soil creep on the nose. The large negative balance obtained on the nose ($-207.0 \times 10^{-3}$ m$^3$/yr) was in contrast to the small positive storage on the side-slope ($+25.1 \times 10^{-3}$ m$^3$/yr) and to the large positive balance in the hollow ($+172.9 \times 10^{-3}$ m$^3$/yr). The nose was the only slope showing a negative balance. Thus, sediment infilled by soil creep into the hollow is primarily from the nose.

4. Slope stability

4.1. Analytical procedure

The slope stability on the nose, side-slope and hollow was analyzed by the infinite slope model. The factor of safety $F_s$ is given by

$$F = \frac{C + \gamma_s(D - H) + \gamma_s \tan \phi}{\gamma_s \cos \beta \sin \beta}$$

where $C$ is the cohesion, $\phi$ is the angle of internal friction, $\gamma_s$ is the saturated unit weight of soil, $\gamma_i$ is

Table 3
Sediment transport rates by soil creep on each slope

<table>
<thead>
<tr>
<th></th>
<th>$S_{\text{min}}$</th>
<th>$S_{\text{avg}}$</th>
<th>$S_{\text{max}}$</th>
<th>$l$ (m)</th>
<th>$Q_{\text{min}}$</th>
<th>$Q_{\text{avg}}$</th>
<th>$Q_{\text{max}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nose</td>
<td>1.5</td>
<td>4.2</td>
<td>7.6</td>
<td>492</td>
<td>75.3</td>
<td>207.0</td>
<td>376.3</td>
</tr>
<tr>
<td>Side-slope</td>
<td>1.2</td>
<td>3.4</td>
<td>6.2</td>
<td>466</td>
<td>58.0</td>
<td>159.5</td>
<td>290.0</td>
</tr>
<tr>
<td>Hollow</td>
<td>0.6</td>
<td>1.6</td>
<td>3.0</td>
<td>55</td>
<td>3.3</td>
<td>9.0</td>
<td>16.4</td>
</tr>
</tbody>
</table>

$S$ = sediment transport rate per unit width. $l$ = perimeter length. $Q$ = sediment transport rate.
the wet unit weight of soil, $\gamma_w$ is the unit weight of water, $D$ is the soil thickness, $\beta$ is the slope angle and $H$ is the water depth in soil. Setting $F_s = 1$ leads to the critical water depth $H_{cr}$ (Iida, 1993) as

$$H_{cr} = \frac{C - \gamma_w \cos^2 \beta (\tan \beta - \tan \phi) D}{\cos^2 \beta (\gamma_{sat} - \gamma_t) (\tan \beta - \tan \phi) + \gamma_w + \tan \phi}$$

Slope failure occurs when the water depth $H$ exceeds the critical water depth $H_{cr}$. This means that slope failure cannot occur if $H_{cr} > D$, because $H$ does not exceed the soil thickness $D$ on hillslopes.

Eq. (6) shows that $H_{cr}$ decreases with increasing $D$ when $\phi < \beta$, whereas $H_{cr}$ increases with increasing $D$ when $\phi > \beta$ (Iida, 1993). This indicates that, on a slope where $\phi < \beta$, the potential of slope failure is high, because $D$ tends to increase with time both by sediment accumulation and weathering. On a slope with such high potential the relationship between $H_{cr}$ and $D$ should be strongly negative. Thus, the relationship between $H_{cr}$ and $D$ was tested on the nose, side-slope and hollow.

The geotechnical properties of soils and slope angles were measured in situ or in the laboratory. The geotechnical properties of the soil ($C$, $\phi$, $\gamma_{sat}$ and $\gamma_t$) were assumed to be constant for each soil type. The residual soil and colluvial soil were tested, and the geotechnical properties of regosol (thin soil) were considered to be the same as those of residual soil because the residual soil is simply regosol with increased soil thickness. The angle of internal friction $\phi$ and the cohesion $C$ were obtained from in situ shear vane testing. The testing apparatus was similar to that used by Matsukura and Tanaka (1983). Prior to the test, the soil layer near the soil-bedrock interface was cleaved to make a horizontal plane, and the layer was saturated with water. The vanes were then pushed into the layer. Since almost all the soils were less than 120 cm thick, low normal stresses ($\sigma = 4.9$ to 34.3 kPa) were applied. Further, the vanes were slowly rotated, and the maximum torque was measured during vane-rotation. Shear stress was then calculated from the torque. The testing was carried out at six sites in residual soil and five sites in colluvial soil. Undisturbed core samples of 2000 cm$^3$ each were collected to measure unit weights. The soil was sampled at two sites at depths of 30 cm.
and 60 cm for each soil type. The soil samples were dried at 110°C in an oven for 24 h, then weighed to calculate the dry unit weight. Wet and saturated unit weights were calculated by the following equations.

\[ \gamma_d = \gamma_s + 0.5n\gamma_r \]  
and

\[ \gamma_{sat} = \gamma_d + n\gamma_w \]

where \( \gamma_d \) is the dry unit weight and \( n \) is the porosity, given by \( n = (\gamma_f - \gamma_d)/\gamma_r \). \( \gamma_r \) is the unit weight of soil particles and is calculated here to be 25.5 kN/m\(^3\). The unit weight of each sample was averaged, and this value was used for the stability analysis. The slope angle \( \beta \) was measured by a slope profiler.

The geotechnical properties of residual soil and colluvial soil were almost identical. As shown in Fig. 12, although the angle of internal friction \( \phi \) of the residual soil at 30.1° was slightly smaller than that of colluvial soil at 36.1°, the cohesion \( C \) was scarcely different for each soil type (3.2 and 2.9 kPa, respectively). Also, the saturated unit weight of the residual soil and colluvial soil (17.1 and 18.2 kN/m\(^3\), respectively) were almost identical (Table 4). These values are essentially similar to those obtained from sandy soils (Okimura and Ichikawa, 1985; Selby, 1993).

A significant difference was found in the distribution of slope angles for these three slopes (Fig. 13, Kruskal–Wallis test, \( H = 15.1, df = 2, p < 0.001 \)). The side-slope was the steepest with an average angle of 44°, while those of nose and hollow were 41° and 37°, respectively. Further, the area in which the angle was steeper than 36.1° occupied 100% of the side-slope, whereas only 68% and 60% of nose and hollow, respectively.

### Table 4

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>( \gamma_d ) (kN/m(^3))</th>
<th>( n )</th>
<th>( \gamma_s ) (kN/m(^3))</th>
<th>( \gamma_{sat} ) (kN/m(^3))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Residual soil</td>
<td>30</td>
<td>7.5</td>
<td>0.71</td>
<td>30</td>
</tr>
<tr>
<td></td>
<td>30</td>
<td>7.5</td>
<td>0.71</td>
<td>60</td>
</tr>
<tr>
<td></td>
<td>30</td>
<td>7.5</td>
<td>0.71</td>
<td>60</td>
</tr>
<tr>
<td></td>
<td>Average</td>
<td>11.8</td>
<td>0.54</td>
<td>14.4</td>
</tr>
<tr>
<td>Colluvial soil</td>
<td>30</td>
<td>11.1</td>
<td>0.56</td>
<td>30</td>
</tr>
<tr>
<td></td>
<td>60</td>
<td>16.7</td>
<td>0.34</td>
<td>60</td>
</tr>
<tr>
<td></td>
<td>Average</td>
<td>13.7</td>
<td>0.46</td>
<td>16.0</td>
</tr>
</tbody>
</table>

\( \gamma_d \) = dry unit weight of soil. \( \gamma_s \) = wet unit weight of soil. \( \gamma_{sat} \) = saturated unit weight of soil. 

\( n \) = porosity.

### 4.2. Stability analysis

Substituting the soil thickness \( D \), the geotechnical properties of the soil \( (C, \phi, \gamma_{sat}, \gamma_r) \) and the slope angle \( \beta \) for Eq. (5), \( H_s \) was calculated for 41, 19 and 20 sites on the nose, side-slope and hollow, respectively, in which the soil survey was carried out.

Fig. 12. Shear-test results of residual soil and colluvial soil.

Fig. 13. Frequency distribution of slope angles for each slope.
Fig. 14. Relations between soil thickness $D$ and critical water depth $H_{cr}$. (see Fig. 2). Only the sites where gravelly colluvial soil was formed were exempted from the analysis because $C$ and $\phi$ could not be measured. $D$ values ranging from 0 to 120 cm were applied for each site, because the thicknesses of almost all the soils were less than 120 cm in this basin.

The results show that everywhere on the side-slope, the condition of $\phi < \beta$; namely $H_{cr}$, decreases with increasing $D$ (Fig. 14). Thus, the stability of the side-slope rapidly decreases as $D$ increases with time. In contrast, $H_{cr}$ increases with increasing soil thickness at several sites on nose and hollow. Moreover, the angles of the lines of the side-slope were significantly steeper than those of nose and hollow (Mann–Whitney’s U-test: $z = 2.00$, $p < 0.05$ for side-slope vs. nose; $z = 5.94$, $p < 0.001$ for side-slope vs. hollow). These results suggest that on the side-slope the potential for slope failure was high, whereas the nose and the hollow were stable.

Nevertheless, even on the nose and in the hollow, $H_{cr}$ exceeds $D$ if $D$ is in excess of 120 cm at most sites. Actually, 120 cm thick soils were present on nose and hollow. These facts suggest that although the potential was low, slope failure was possible also on the nose and in the hollow. However, slope failure can occur only when porewater pressure is produced. Subsurface water converges into the hollow, potentially developing a high porewater pressure, while it diverges from the nose (Kirkby and Chorley, 1967; Anderson and Burt, 1978; Whipkey and Kirkby, 1978; Tsukamoto et al., 1982; Iida, 1984; Burt and Butcher, 1985; Wilson and Dietrich, 1987; Terajima and Moroto, 1990; Onda et al., 1992; Fernandes et al., 1994). Thus, slope failure would most likely occur on the side-slope and in the hollow, whereas it would be unlikely to occur on the nose.

5. Landscape evolution due to soil creep and slope failure

Soil creep processes and slope stability show remarkable contrasts between the nose, side-slope and hollow. Sediment transport rates by soil creep clearly demonstrate the significance of denudation on the nose and of infilling in the hollow. Stability analysis shows the high potential of slope failure on the side-slope and in the hollow. The slope failure that would occur on the side-slope infills soil materials into the hollow in addition to that accumulated through soil creep, because the hollow is located below the side-slope. The infilling due to soil creep and slope failure results in instability in the hollow. Thus, slope failure eventually evacuates the hollow, after these processes have developed a thick soil layer in the hollow.

The hollow-infilling processes operating to remove materials on the nose and side-slope are likely to be important in the long-term landscape evolution of the headwater basin. The denudation on the nose is gradual, as shown in the creep rates. The short slope (ca. 10 m) of the side-slope results in small slope failures on it. Thus, the landscape change due to those processes is likely to be gradual. However, evacuation of the resultant infilling materials in the hollow will drastically change the landscape of the
zero order basin, and resultant debris flows will also affect the landscape of the head water river (Hack and Goodlett, 1960; Dietrich and Dunne, 1978; Lehre, 1982; Tsukamoto et al., 1982; Reneau and Dietrich, 1987; Crozier et al., 1990; Montgomery et al., 1991). Although the time required for a cycle of infilling and evacuation was not derived in this basin, dated colluvial soils in headwater basins in humid temperate climates show that the time ranges from 4000 to 15,000 yr (Dietrich and Dorn, 1984; Marron, 1985; Reneau et al., 1986, 1989; Reneau and Dietrich, 1990, 1991; Yoshinaga and Koiwa, 1996). Thus the hollow infilling processes, which cause only gradual change in the short term will cause drastic change in thousands or tens of thousands of years.

6. Conclusions

This study revealed that soil creep removes soil materials on the nose and infills them into the hollow, whereas slope failure removes materials on the side-slope and infills them into the hollow. When these processes develop a thick soil in the hollow, slope failure evacuates the hollow. These results show the significance of soil creep on the nose and of slope failure on the side-slope and in the hollow in the landscape evolution of the zero order basin. Although the landscape change due to soil creep on the nose and slope failure on the side-slope would be gradual, the resultant infilling greatly affects the slope stability in the hollow in thousands or tens of thousands of years. Thus, it is concluded that hollow infilling processes operating to remove materials from the nose and side-slope play the most important role in the long-term landscape evolution of headwater basins.

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References


