Abstract

In Japan, Typhoon Talas (T1112) induced many rapid deep-seated landslides in the Kii Peninsula. A landslide is one of main processes associated with debris flow initiation. In particular, deep-seated catastrophic landslides can lead to large-scale debris flows that seriously impact human welfare. Previous studies have explored the roles played by geology and geological structure. However, no single, widely used physical model is employed to analyze such landslides. Here, we focus on slope scale and gradient, and explore the relationship between height and gradient for several slopes. We found that the height of slopes exhibiting various gradients was limited, where the greater the gradient, the lower the height; the relationship was amenable to slope stability analysis. We developed a physical model of deep-seated landslides and identified regions at risk. We focused on slopes where typhoon Talas caused such landslides and used detailed topological (LiDAR) data collected before and after the typhoon to measure slope gradients and relative heights. Our model effectively localized deep-seated landslides, although we assumed that the strength of weathered rock was uniform throughout the study area, based on data on the side-slope gradients and relative heights of land abutting the Totsugawa River.

Keywords: deep-seated catastrophic landslides, slope stability analysis, Typhoon Talas, Kii peninsula

1. Introduction

In Japan, Typhoon Talas caused many large-scale, deep-seated catastrophic landslides in 2011, principally in the Kii Peninsula. The Shimanto Belt (a paleo-accretionary prism) is widely distributed throughout the Kii Peninsula and is well known for its deep-seated landslide events. In addition, many of the deep-seated landslides caused by Typhoon Talas occurred in “dip slope” formations; this type of geological structure is considered to be a precursor to that responsible for deep-seated landslide occurrences.

Shallow landslides have been physically modeled by combining slope stability analysis with a model of underground water flow; many previous studies have evaluated landslide risks (e.g., Montgomery and Dietrich, 1994). However, only a few models of deep-seated catastrophic landslides are available, although slope risk evaluation methods based on geological structure and small-scale slope deformations have been suggested to be important in this context.

Schmidt and Montgomery (1995) focused on slope size, height, and inclination, and explored the relationship between the latter two variables. For a particular inclination, slope height is limited, where the steeper the slope, the smaller the limit; this relationship is amenable to slope stability analysis. Matsushi et al. (2014) used detailed topographical data collected pre- and post-landslide to estimate soil parameters; it was possible to explain landslides by reference to height limits for slopes of various inclinations, as proposed by Schmidt and Montgomery (1995). Korup and Schlunegger (2007) similarly analyzed bedrock collapse in the Swiss Alps. Previously, we assumed that widening of water channels caused by erosion when landslides were triggered by dam overflows reflected repetitive side-bank collapse triggered by changes in the soil mechanical balance attributable to riverbed erosion; we
formalized this balance by reference to the internal soil stress and strength (Yoshino et al., 2013). Thus, we explored changes in soil mechanical balance when the relative height of a slope increased because the riverbed was lowered by erosion, using the method of Schmidt and Montgomery (1995). However, other studies considered that deep-seated catastrophic landslides occurring near rivers were attributable to loss of soil and rock mechanical balance when the relative height was increased by elevation of the slope (e.g., Matsukura, 1987a and 1987b). On the one hand, landslides caused by dam overflows trigger erosion that widens water channels, while on the other hand deep-seated catastrophic landslides create valleys. The processes involved differ in terms of both genesis (riverbed lowering and ridge upheaval, respectively) and speed. However, if Matsukura (1987a and 1987b) is correct, both processes could be described using a model similar to that of Yoshino et al. (2013). Therefore, we explored the relative heights of slopes that participated in deep-seated catastrophic landslides after typhoon Talas (2011), and of slopes that remained stable. We used detailed topographical data collected both before and after the event. We then determined whether the side-bank collapse model that we had earlier developed could be realistically applied to typhoon-triggered landslides. We suggest that our approach can be used to predict the location, scale, and shape of deep-seated catastrophic landslides, and the challenges they pose.

2. Methods

2.1. Working hypothesis

Previous works on the development of mountain topography considered that the rates of upheaval and erosion of channels with large drainage areas were generally uniform, thus varying little with elevation (e.g., Kaizuka, 1978). However, in mountain streams with shallow slopes and small drainage areas, the rate of erosion is lower than the upheaval rate, and elevation will thus increase. Specifically, for ridges, erosion caused by running water is zero. If the valley and ridge positions are considered to remain constant, the valley elevation does not change but both the slope relative height and inclination increase over time. Yoshino et al. (2013) formalized the collapse of water channel side-banks, which sometimes occurs when landslides caused by overflowing dams trigger erosion. The analysis was based on the relationship between slope inclination and the relative height limit developed by Schmidt and Montgomery (1995). Although the time scales of landslides triggered by dam overflow erosion and mountain topographic development vary greatly, they share the feature that the valley and ridge, corresponding to the side-bank shoulder of a landslide caused by dam overflow, do not change. Both the inclination and relative slope height, i.e., the slope and relative height of the side-bank, increase with time, and the slope (side-bank) ultimately becomes physically unstable, collapsing when the relationship between the relative height and inclination reaches a critical point. Therefore, we assumed that, considering that the horizontal positions of the valleys and ridges does not change, deep-seated catastrophic landslides are more likely to occur in a slope that became unstable due to a time-dependent elevation of the ridge, even though the elevation of the valley remains constant. We verified this hypothesis using the concept of water channel side-bank collapse, as caused by erosion when dam overflow triggers a landslide [Yoshino et al. (2013)]. It can be thought that the underground structures, like bedding angle and so on, should be affected heterogeneity of bedrock strength and groundwater movement. However, in this study, we did not consider heterogeneity of underground condition. So here we tested effects of topography on deep-seated catastrophic landslide occurrence.

2.2. The model

Figure 1 shows a schematic diagram of the landslides studied. The model assumes that the relative slope height ($H$) increases because of slope upheaval and valley bottom erosion, while $L$ remains constant; when the slope becomes unstable, it collapses over width $D$, creating a wedge-shaped landslide front of inclination $\beta$, thus increasing the slope width ($L$). The slope safety ratio is derived via stability analysis:

$$F_s = \frac{C_i + (W_i \cos \beta - U_i) \tan \phi}{W_i \sin \beta}$$

(1)

Here, $W_i = H_i D, \gamma / 2, C_i = c \left((L_i + D_i)^2 + H_i^2\right)^{1/2}, U_i = u \left((L_i + D_i)^2 + H_i^2\right)^{1/2}, \beta_i = \arctan\left(\frac{H_i}{(L_i + D_i)}\right)$, where $c, \phi, \gamma$ and $u$ are the adhesiveness of soil, internal friction angle, unit weight, and pore water pressure, respectively. The subscript
When $H=H_{C0}$, the first landslide occurs, causing $\beta_0$ to develop. When $H=H_{C1}$, collapses with the width $D_1$ and a wedge-shaped landslide surface with inclination $\beta_1$ develops. Thus, the slope shape after the first landslide is shown in Figure 1(b)', and $L_{i+1}$ can be expressed as:

$$L_{i+1} = L_i + D_i$$ (6)

We assume that, as long as the slope continues to satisfy the stable conditions ($F_S > 1$) of Eq. (1), $L_i$ does not change; only $H$ increases. The relationship between $\theta$ and $H$ is:

$$\tan \theta = \frac{H}{L_i}$$ (7)

When $L_0$, $c$, $\phi$, and $\gamma$ are defined as above, the relationship between $H$ and $\theta(H)$ can be calculated. Figure 2 shows the process in graphical form. If the slope relative height, $H$, increases because of slope upheaval, $\theta$ also increases when $L_i$ is held constant (thick solid line). Later, when $H$ attains $H_{C0}$, the first landslide occurs, causing $\theta$ to decrease as shown in Figure 1. Later, if upheaval continues, $H$ and $\theta$ increase once more [Eq. (7)], and a second landslide develops similarly. We assume that the width of the riverbed does not change either when the relative height of the slope increases or when the slope collapses.
3. Sites and methodology

3.1. Deep-seated catastrophic landslide sites

We selected three (among many) deep-seated catastrophic landslides that occurred in the Kii Peninsula after typhoon Talas in 2011, i.e., those at Akadani, Akadani-higashi, and Nagatono (Figure 3). All three slopes have “dip slope” type configurations.

3.2. Data

We used 1-m digital elevation model (DEM) data collected by airborne lasers. The pre- and post-landslide data were obtained in 2009 and between December 2011 and February 2012, respectively. For Akadani, a few re-slides and landslide expansions occurred after the deep-seated catastrophic event (Sakurai et al., 2015); the data did not encompass these later events.

3.3. Analysis

We evaluated areas to 500 m up- and down-stream from each landslide, as follows. First, using pre-landslide data, we set reference points at 20-m intervals along the ridges. Then, we drew lines from these points toward the points in the region of the slope toes (where the slope bottoms and the flats between the banks intersect) where the inclinations were steepest. We then divided the cross-sections into those associated with landslides (termed CSLs below) and not associated with landslides (termed NCSLs below) (Figure 4). When at least 66% of the lines were included in the landslide, the cross-sections were considered to be CSLs, and the relationship between relative height ($H$) and slope inclination ($\theta$), from the slope foot to the ridge, was examined with respect to all lines. Then, ground strength parameters ($c$, $\phi$, and $\gamma$) were repeatedly varied within specific ranges ($c$: 5 to 300 kN/m$^2$, $\phi$: 10 to 40$^\circ$, $\gamma$: 15 to 25 kN/m$^3$) and entered into Eq. (2) to optimize landslide prediction by $H_c$ (the percentage of cross-sections correctly identified as landslide sections among the actual CSLs) and the cover rate (the percentage of actual CSLs among cross-sections with $F_S < 1$). We then derived the combination of parameters that optimally predicted landslide occurrence/non-occurrence.
Although many landslides commenced from the slope foot, the entire slope did not necessarily collapse. Thus, using pre-landslide data, we set measurement points every 10 m along each line running horizontally from the slope foot, and explored the relationship between the relative height and inclination according to those of the slope foot (Figure 5). We thus extracted unstable regions in the middle of slopes, and not only in the ridges; this was especially important when the slope was convex. Next, using the post-landslide data, we performed a similar analysis to determine whether the landslides had eliminated unstable slopes. If our hypothesis was correct, it would be possible to calculate the decrease in slope inclination caused by a landslide (arrows in Figure 2), in turn allowing calculation of the size and shape of the wedge-shaped landslide front. We compared the shape before and after a landslide; we now show how to estimate the shape and the front size.

4. Result

4.1. Longitudinal slope shapes

The longitudinal lines drawn in the three landslide sites are shown in Figure 6. Figure 7 shows the shape of the principal longitudinal line of each slope, i.e., that located in the central CSL. In the longitudinal direction, the slope of the Akadani-higashi terrain prior to the landslide [Fig. 7(b)] is linear or (downward) convex. However, the pre-landslide terrains of Akadani [Fig. 7(a)] and Nagatono [Fig. 7(c)] are steepest from the bottom to the middle of the slope, gradually becoming gentler from the middle to the top, and trending (upward) convex. In Figure 7, the Akadani-higashi trend differs from those of Akadani and Nagatono, attributable to the fact that the 2011 landslide at Akadani-higashi was a re-slide of the internal region (at the top of the slope) of the landslide produced by the major Totsugawa flood of 1889.

4.2. Relationship between slope relative height and inclination

Figure 8 shows the pre-landslide relationship between the relative slope height \( H \) and inclination \( \tan \theta \) (from the foot to the ridge) relative to the longitudinal lines drawn at the three sites (Fig. 6). In the graph, CSLs and NCSLs are represented by + and ○, respectively. At Akadani [Fig. 8(a)] and Nagatono [Fig. 8(c)], the NCSLs downstream from the landslide, i.e., those within the dashed frames were in areas of low relative height but with large inclinations; the upstream NCSLs were in areas of large relative height and small inclinations. For Akadani-higashi [Fig. 8(b)], the CSLs were in areas of large relative height and pronounced inclination. Thus, although the Nagatono data are not as clear as those from the two other sites, most CSLs are in the upper right corner of the graphs, i.e., in areas with large relative heights and slope inclinations). Also, as indicated in Table 1 and Figure 8, the calculated ground strength parameters were relatively high \( (\phi: 10 \text{ to } 15^\circ, \gamma: 18 \text{ to } 22 \text{ kN/m}^3, c: \geq 200 \text{ kN/m}^2 \text{ in all cases}) \), and the sites were similar. There are differences in best-fitted ground strength parameters although these three landslides underlain by same bedrock geology. While, Schmidt and Montgomery (1995), evaluated ground strength in the Chuckanut Formation in the northern United States as \( \phi \) of 17 to 21° and \( c \) of 120 to 150 kN/m².
using similar method. The degree of variation of this study is similar to that of Chuckanut Formation, suggesting that ground strength controls landslide occurrence might be varied in space even in the same bedrock type.

5. Discussion

5.1. Estimation of landslide slope

As shown in Section 4.2, in most upstream and downstream slopes, landslides developed in the upper right corners of Figure 8, i.e., in areas with large relative height and slope inclination. Hence, our method identifies slopes located along high-risk mountain streams that are at relatively high risk of landslides. It was possible to distinguish landslide from non-landslide slopes by the bedrock strength. In Akadani, one point in an area of large relative height slightly exceeded the relationship described by Eq. (2). Akadani noted multiple re-slides and landslide expansions after the landslide caused by typhoon Talas but, as our data were obtained prior to these events, and as an expanding landslide later developed in the upper region of the landslide site, it is possible that this was a component of the
5.2. Estimation of landslide scale and shape

Here, we discuss landslide scale in terms of the relationships between the real and calculated data. Figure 9 shows the relationship between the relative height and inclination of the principal slope cross-section, calculated every 20 m horizontally from the slope foot; we used both pre- and post-landslide data in this analysis. Except for a few points with steep inclinations in areas of low relative height, the pre-landslide Akadani inclination was maximally $\tan \theta = 0.79$ ($H = 380$ m), but became relatively uniform after the landslide, ranging from $H = 200$–$600$ m with $\tan \theta = 0.6$. Similarly, at Akadani-higashi, the post-landslide inclination became near-uniform at $H \geq 300$ m and $\tan \theta = 0.6$. At Nagatono, the maximum inclination pre-landslide was $\tan \theta = 0.75$ ($H = 200$ m), but reduced markedly post-landslide, to $\tan \theta = 0.56$ ($H = 180$ m); the relative height increased, as did the slope inclination. The shape differences post-landslide are evident in Figure 7; at Akadani and Nagatono, the inclinations fell particularly sharply. However, the actual decreases were about 40–60% of the calculated values. In other words, the actual falls were smaller than those calculated. The region that we analyzed experienced multiple deep-seated catastrophic landslides after the heavy rains of 2011; many originated on northwest-oriented “dip slope-type” slopes, indicating that geological structure plays a major role in the development of such landslides (Hiraishi and Chigira, 2011). Thus, it is possible that landslides develop over structurally soft surfaces, such as bedding planes. In addition, recent studies suggest that such soft surfaces may have been earlier deformed by the long-term effects of gravity; the extent of deformation may determine the landslide scale (Chigira et al., 2013). It is also possible that such geological structures are associated with high local pore water pressures (Jitozono et al., 2004). Our model assumes that geological strength is uniform and does not consider underground water. This may be why the model did not accurately predict post-landslide decreases in inclination; thus, further refinement is required.
6. Summary and future tasks

In areas with similar geological structure and soil strength, our model successfully reproduced the relationship between the side slopes and their relative height along the course of the Totsugawa River. This does not contradict the hypothesis wherein, considering that the horizontal positions of the valleys and ridges does not change, deep-seated catastrophic landslides are more likely to occur in a slope that became unstable due to the time-dependent elevation of the ridge, even though the elevation of valley remains constant. Thus, we used measurable physical parameters to identify slopes that might trigger deep-seated catastrophic landslides. However, when we explored whether the model estimated landslide shape and scale in slope cross-sections, we found that, for Akadani and Nagatono, although inclinations were significantly reduced after the landslides, the extent of the reduction was less than predicted.

We modeled slope shape and landslide formation in a relatively simple manner. However, although many landslides commence at the slope foot, the entire slope does not necessarily collapse. Therefore, when our model is further applied to actual phenomena, such as landslides developing in the mid-sections of uniform or concave slopes, inconsistencies are to be expected.

In this study, we only focused on dip-slope. Thus, to use this method as "a predictive tool in other locations or at a larger scale", it would be necessary to consider roles of underground characteristics, such as underground structure and spatial variability of ground strength controls landslide occurrence. Thus, further work is needed. In future, we will analyze slope topographical characteristics in greater detail and underground condition and evaluate more actual landslide sites; we will also consider terrain changes caused by the Totsugawa disaster of the Meiji era.

References

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