Long-Runout-Landslide-Induced Debris Flow: The Role of Fine Sediment Deposition Processes in Debris Flow Propagation

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Abstract Landslide-induced debris flows can travel long distances. Many field studies, laboratory experiments, and theoretical studies have been conducted to clarify the mechanisms of long-runout landslides. However, information concerning landslide-induced debris flows is still inadequate despite several explanations proposed. Thus, we collected various forms of data, including video images and light detection and ranging data, for a landslide-induced debris flow that occurred on 28 July 2015, in Fukaminato River, Japan, and clarified the debris flow behavior from initiation to deposition. We showed that the water content of the debris flow should only be approximately 30% immediately after landslides occurred at the site. However, a fluid phase-dominated flow flowed over 1.2 km, expanding the damaged area. We inferred that the fine sediments behaved as part of the interstitial fluid in the debris flow and that, during debris flow deposition, some fine sediments were stored in the interstitial space of the riverbed and did not contribute to forming the skeleton of deposits. We conducted a numerical simulation to test the processes of fluid phase-dominated flow propagation, focusing on the flow and deposition of fine sediments. Our simulation results suggest that fine sediments’ flow and depositional processes significantly influence both the initiation of fluid phase-dominated debris flow and the travel distance of landslide-induced debris flow.

Plain Language Summary Landslides can cause debris flows that travel long distances. Although several studies have been conducted to clarify the mechanisms of long-runout landslides, those based on a series of data from the initiation to the deposition of the debris flow are limited. This is because landslides often occur without any precursor signals. We collected various forms of data, including video images and light detection and ranging data, for a landslide-induced debris flow that occurred on 28 July 2015, in Fukaminato River, Japan. We then clarified the debris flow behavior from initiation to deposition. Furthermore, debris flow deposition was clarified using a debris flow numerical simulation. At the site, debris flow containing sediments of various grain sizes flowed upstream. A fluid phase-dominated debris flow containing a large amount of fine sediment flowed downstream. We suggest that, during debris flow deposition, some fine sediments formed a skeleton structure, whereas other sediments were stored within the void space of the skeletal sediments. Our results suggest that the depositional processes of fine sediments may cause fluid phase-dominated and long-traveling-distance debris flow.

1. Introduction

Landslide-induced debris flows can travel long distances and seriously damage infrastructures (Iverson & George, 2014; Tai et al., 2019). Therefore, predicting areas most at risk of future landslides is essential in mitigating the effects of such disasters. In the last decades, landslide mobility has been widely evaluated using relationships between the travel distance of landslides (L) and the height between the landslide scar and the lower end of the deposited area (H) (e.g., Hsü, 1975; Legros, 2002; Scheidegger, 1973). The value of H/L has been revealed to be related to the landslide scale (i.e., the area or volume of the landslide) in various environments, not only on Earth but also on other planets (e.g., Crosta et al., 2018; Iverson, 1997; Legros, 2002; Lucas et al., 2014). These empirical scaling relationships are important in predicting areas affected by landslides and debris flows (e.g., Hürlimann et al., 2008).

However, the travel distance of a landslide is sometimes surprisingly long compared with predictions based on such empirical relationships (e.g., Iverson et al., 2015; Usuki et al., 2006). Thus, many field studies, laboratory experiments, and theoretical studies have been conducted to clarify the mechanisms of long-runout landslides,
and several explanations have been proposed (e.g., Hungr & Evans, 2004; Iverson, 1997, 2015; Johnson et al., 2016; Nishiguchi et al., 2011; Sidle & Chigira, 2004; Wang & Sassa, 2003). Several studies reported that the high mobility of landslides results from the liquefaction of water-saturated sediments at their base (e.g., Iverson et al., 2015; Wang & Sassa, 2003). Moreover, landslides sometimes transform entirely into debris flows and travel long distances. Further, several recent studies showed that debris flows grow in size as they entrain materials from their beds (e.g., de Haas et al., 2020; Iverson et al., 2011). Such landslide-induced debris flows seriously endanger human life (e.g., Nishiguchi et al., 2012; Sidle & Chigira, 2004). In this study, we focused on landslide-induced debris flows.

At present, field information concerning landslide-induced debris flow is inadequate. For example, although the water content of landslides should influence the fluidity of both landslides and debris flows (e.g., Iverson et al., 2015), information on water content is limited. This is because landslides often occur without any precursor signals. Direct observation data on runoff propagation (e.g., flow velocity and depth and debris flow types) are also limited, with a few exceptions (e.g., Sosio & Crosta, 2009). Consequently, in this study, we gathered new data on unique landslide-induced debris flows in Fukaminato River in Kagoshima, Japan, which occurred on 28 July 2015. This landslide occurred during a rain-free period. All processes, including landslide occurrence, the transition to debris flow, and the flow and stop of the debris flow, were recorded on video. Moreover, light detection and ranging (LiDAR) data for before and after the landslide are available. The first objective of this study is thus to clarify the entire process of landslide-induced debris flows, such as the initial state of the landslide, the transition from landslide to debris flow, and the debris flow runout and deposition, using detailed field data.

In the last few decades, several numerical simulation models have been applied in describing landslide-induced debris flows (e.g., Egashira et al., 2016; Nishiguchi et al., 2011, 2014; Tai et al., 2019; Uchida et al., 2013). In these simulations, researchers described the large-scale fluidity of landslide-induced debris flows. However, the fluidity of sediment-rich debris flows should be small because the increase in collisions between particles increases energy dissipation (e.g., Hürlimann et al., 2015). Thus, previous modeling efforts often set the friction coefficient of solid materials to be relatively small and/or the fluid density of the interstitial fluid to be large to describe large-scale fluidity (e.g., Egashira et al., 2016; Iverson & George, 2016; Nishiguchi et al., 2011; Tai et al., 2019; Uchida et al., 2013). Moreover, various studies revealed that the increase in the fines fraction enhanced the fluidity in stony debris flows (e.g., D’Agostino et al., 2010; de Haas et al., 2015; Hürlimann et al., 2015; Iverson et al., 2010; Kaitna et al., 2016; Sakai et al., 2019; Sosio et al., 2007; Uchida et al., 2021). It has been considered that fine sediments moved as turbulent flows with interstitial water (e.g., Iverson, 1997). These fine sediments thus helped enhance buoyancy forces acting on solid phases (e.g., Iverson, 1997; Iverson & Denlinger, 2001; Pitman & Le, 2005; Tai et al., 2019) and reduce energy dissipation due to decreased collisions between particles (e.g., Iverson & George, 2014) and drag stress between the fluid and solid phases (e.g., Pitman & Le, 2005). Recently, the effects of fine sediments have been included in numerical simulations and applied to describe long-runout-landslide-induced debris flows (e.g., Egashira et al., 2016; Nishiguchi et al., 2011, 2014; Tai et al., 2019; Uchida et al., 2013).

However, the depositional process of fine sediments has not been fully characterized in previous models. Takahashi et al. (2001) and Uchida et al. (2013) argued that there are two primary possibilities for the depositional process. First, the fine sediments in debris flows could behave as solids immediately before deposition. Second, they could behave as fluids and flow with the interstitial fluid until after deposition. Previous studies did not test the roles of depositional processes on travel distance. Consequently, the second objective of our study was to examine the influence of the depositional processes of fine sediments on debris flow propagation on the basis of numerical experiments at Fukaminato River using parameters derived from detailed field surveys.

2. Hypothesis

2.1. Phase-Shift of Fine Sediments

Stony (boulder-rich) debris flows have been conceptualized as having two phases: the fluid and solid phases (e.g., Egashira et al., 2016; Iverson, 1997; Takahashi, 1991). Previous studies considered that relatively coarse sediments in debris flow move in a laminar manner (e.g., Iverson, 1997). In contrast, relatively fine sediments exhibit turbulent flows with interstitial water. Processes involving behavioral transitions from part of the solid phase to part of the fluid phase of the debris flow are termed “liquefaction” (Hotta et al., 2013); “phase shifting” (Egashira et al., 2016); and “phase transition” (Egashira et al., 2016).
et al., 2016), “phase-shift” (Uchida et al., 2021), and “hydrodynamic suspension” (Iverson & George, 2014). Fine sediments behaving as fluid phases are called “suspended sediment” (Kaitna et al., 2016), “fluid phase sediment” (Sakai et al., 2019), “phase-shifted sediment” (Uchida et al., 2021), and so on. Although various terms are used for this phenomenon, we will call it phase-shift, following the research that has used the same numerical simulation model as that of this research. Fine sediments that behave as fluid phases with interstitial fluids are called “phase-shifted sediments,” whereas coarse sediments that act as solid phases in the debris flow are termed “non-phase-shifted sediments” (Figure 1a). The occurrence of the phase-shift phenomena has been verified in flume experiments (e.g., Hotta et al., 2013; Sakai et al., 2019).

Thus, we assumed that the sediment concentration for all sediments ($C_s$; i.e., coarse and fine sediments) could be described by the coarse sediment ($C_c$) and fine sediment ($C_f$) concentrations, as follows:

$$C_s = C_c + C_f$$ \hspace{1cm} (1)

The relation between $C_s$ and the volumetric ratio of water in the debris flow ($C_w$) can be described as

$$C_s + C_w = 1$$ \hspace{1cm} (2)

2.2. Deposition of the Phase-Shifted Sediment

Takahashi et al. (2001) classified debris-flow-deposited sediments into two groups. The first includes sediments that form a skeleton structure, and the other includes sediments stored within the void spaces of the skeletal sediments. In this study, we termed the first group “skeleton-forming sediments” and the second “void-space-stored sediment.” Thus, the sediment concentration of deposits on the riverbed ($C_{sd}$) can be described as

$$C_{sd} = C_{sdv} + C_{sdv}$$ \hspace{1cm} (3)

where $C_{sdv}$ and $C_{sdv}$ are the concentrations of the skeleton-forming and void-space-stored sediments in deposits, respectively.
Takahashi et al. (2001) assumed that non-phase-shifted sediments form the skeleton of deposits, whereas phase-shifted sediments are stored in void spaces when the ratio of the fine particles in the bed is smaller than 26%. Uchida et al. (2013) instead thought that all sediments, including phase-shifted sediments, are deposited as the skeleton of deposits because the pore scale for interstitial fluids would likely decrease immediately before deposition. Major and Iverson (1999) and Major (2000) examined the depositional processes of sediments in debris flows through laboratory experiments. According to their experiments, the total fluid pressure in the deposited materials was equal to the unit weight of the debris flow immediately after deposition. This suggested that the sediments still existed as fluid phases and did not form a skeleton. Interstitial fluids then dissipated through Darcian seepage, allowing buoyant grains to settle and form the skeleton. This experimental result agrees with the hypothesis proposed by Uchida et al. (2013), indicating that phase-shifted sediments might contribute to the formation of the skeleton of deposits.

Some phase-shifted sediments are converted to the solid phase immediately before deposition when the turbulence magnitude of interstitial fluids becomes small due to the decrease of pore spaces. Conversely, fine particles in phase-shifted sediments might keep their phase-shifted state if the deposition duration is short compared with the timescale for grain settling in static water. The laboratory experiments of Major and Iverson (1999) and Major (2000) were conducted under static conditions. Therefore, fine particles took long to be deposited, resulting in all sediments depositing as the solid phase. We can therefore hypothesize that, during the deposition of debris flows, phase-shifted sediments are classified into two: the first continuing in the phase-shift state and the other converting into the solid phase (Figure 1b). This hypothesis is similar to the deposition process argued by the sedimentological studies of fluvial deposits (e.g., Frings et al., 2008). For example, Frings et al. (2008) reported two types of sediments in riverbed material on the Rhine River in Germany and the Netherlands: the bed-structure sediment (i.e., skeleton-forming sediment) and the pore-filling sediment (i.e., void-space-stored sediment).

In this study, we considered that, after deposition, the skeleton of deposits, as in Takahashi et al. (2001), and that phase-shifted sediments could form void-space-stored sediments (Figure 1c). Thus, the skeleton-forming sediment concentration ($C_{sdc}$) can be described as follows:

$$C_{sdc} = C_{dsdc} + C_{dsf}$$

where $C_{dsdc}$ and $C_{dsf}$ are the volumetric concentrations of deposited sediments derived from non-phase-shifted sediments and of the sediments converted from the fluid phase to the solid phase, respectively.

Here, we defined the ratio of the phase-shifted sediment volume during deposition to the phase-shifted sediment volume just before deposition as “$k$.” Consequently, when $k$ is 0, all phase-shifted sediments in the debris flow are converted to the solid phase during deposition. When $k$ is 1, all phase-shifted sediments keep their phase-shift state during deposition. de Haas et al. (2015) reported different deposition behaviors for clay and sand, suggesting that the value of $k$ might be controlled by sediment type. Moreover, previous studies indicated that the deposition of fine sediments is strongly affected by the shape of the particle size distribution of the deposited material (de Haas et al., 2015; Frings et al., 2008; Major, 2000). These studies suggested that the value of $k$ may largely depend on the shape of the particle size distribution of the deposited material, suggesting that the value of $k$ might vary in time and space. However, we considered that the value and characteristics of $k$ have not been sufficiently understood. Thus, in this study, $k$ is first simply assumed to be constant in time and space. $C_{dsdc}$ and $C_{dsf}$ can then be described as follows:

$$C_{dsdc} = C_{dsb} \times \frac{C_v}{(1 - k)C_f + C_v}$$

$$C_{dsf} = C_{dsb} \times \frac{(1 - k)C_f}{(1 - k)C_f + C_v}$$

Following Major and Iverson (1999) and Major (2000), we assumed that excess interstitial fluid should return to the flow through Darcian seepage. Thus, the volumetric ratio of the sediments stored in void spaces to those in pore spaces equals that of the phase-shifted sediment volume during the deposition of interstitial fluid. This indicates that the void space sediment concentration in deposits ($C_{dv}$) can be described as follows:

$$C_{dv} = (1 - C_{dsb}) \times \frac{kC_f}{kC_f + C_w}$$
Consequently, three mass balance equations for the entire flow, coarse sediments, and fine sediments can be derived, respectively, as follows.

\[
B \frac{\partial h}{\partial t} + \frac{\partial uhB}{\partial x} = iB \tag{8}
\]

\[
B \frac{\partial Cc}{\partial t} + \frac{\partial CcuhB}{\partial x} = iBC_{vdc} \tag{9}
\]

\[
B \frac{\partial Cl}{\partial t} + \frac{\partial CluhB}{\partial x} = iB (C_{edc} + C_{edv}) \tag{10}
\]

where \( h \) (m) is the flow depth, \( u \) (m/s) is the flow velocity, \( B \) (m) is the flow width, \( x \) (m) is the horizontal distance, \( t \) (s) is time, and \( i \) represents the erosion and deposition rate.

From Equations 5–7, 9 and 10 can be respectively converted to

\[
B \frac{\partial h}{\partial t} + \frac{\partial uhB}{\partial x} = iB \frac{C_c}{C_c + (1-k)C_l} \tag{11}
\]

\[
B \frac{\partial Cc}{\partial t} + \frac{\partial CcuhB}{\partial x} = iB \left\{ \frac{C_{edc}}{C_c + (1-k)C_l} + \frac{(1-k)C_l}{kC_l + C_v} \right\} \tag{12}
\]

3. Study Site

Our study was conducted in the catchment area of Fukaminato River, Kagoshima Prefecture, Kyushu, Japan (31.6°N, 130.8°E; Figure 2a). The catchment is located in the Aira caldera wall, facing Kinko Bay. The approximately 20 × 20-km Aira caldera was formed by a series of large-scale pyroclastic eruptions around 22,000 years ago (Aramaki, 1984).

Fukaminato River has an approximately 6.4-km² catchment (Figure 2b). The catchment mainly consists of topographic plateaus and valleys. The valleys cut into the plateaus and form steep hillslopes at their edges. The plateaus are relatively gentle (4°–9°) and have elevations ranging from around 200 to 450 m. The heights of the steep hillslopes at the edges of a plateau range from a few tens of meters to approximately 150 m. At the toe of the steep hillslopes, taluses form because of landslides and rockfalls. The longitudinal gradient of the valley studied is between 3° and 17°. The mean annual precipitation and mean temperature are 2,758 mm and 16°C, respectively, at the Kihoku meteorological observation station (approximately 5 km from the landslide site). The tops of the plateaus and steep hillslopes are mainly covered by evergreen or coniferous forest. Some plateaus (and valley bottoms) are used for residential or cultivation purposes.

The plateaus are underlain by sedimentary rocks of the Cretaceous to Paleogene Shimanto Group. Tuffaceous sedimentary rocks, Neogene welded tuff, Quaternary sand and gravel layers, and deposited volcanic soils lie on the Shimanto Group (Ayaori, 2017; Honda, 2016). Borehole survey was conducted on the plateau, the location of which is shown in Figure 2c. This showed that the thicknesses of the deposited volcanic soils, sand and gravel layers, welded tuff, and tuffaceous sedimentary rocks were around 9, 13, 15, and 6 m, respectively. The borehole survey also suggested that the interfaces between each unit are almost flat.

Six landslides occurred on one of the steep hillslopes at the plateau’s edge in June and July 2015 (Figure 2c). These landslides were debris flows based on the landslide classification proposed by Hungr et al. (2014). On June 24 and 5 July 2015, two landslides occurred and induced debris flows. These debris flows impacted a national road around 1 km downstream (Figure 2c). Consequently, on 24 June 2015, the central and local (Kagoshima prefectural) governments installed four Video Cameras to develop an early-warning system for disasters arising from secondary landslides. The locations of these Video Cameras are shown in Figure 2c (Ayaori, 2017). On 28 July 2015, two more landslides occurred on the same hillslope. The first landslide occurred at 12:42LT, and the second occurred at 12:48LT. There was no rain on July 28. However, the cumulative rainfall amounts from the previous 5 and 10 days were approximately 48.5 and 274.5 mm, respectively. The monthly total precipitation values in June and July in 2015 were 1,311 and 687 mm, respectively, 2.4 times and 1.7 times larger than the mean monthly precipitation values for June and July from 1981 to 2010. We can therefore consider the
groundwater level in the plateau to be high; this may have contributed to the frequent landslide initiation. In this study, we focused on the last two landslides because the Video Camera captured the entire process, including landslide occurrence, the landslide to debris flow transition, debris flow runout, and deposition.

The slope gradient of the studied steep hillslope was around 45° before the last two landslides occurred. The elevations of the top and bottom of a slope were around 200 and 140 m, respectively (Figure 2b). The lateral length from the bottom of the steep hillslope to the end of the talus is around 500 m. The mean surface gradient of the talus is around 15°. The relatively gentle valley floor extends downstream of the talus to the sea with a slope gradient ranging from 0.5° to 3°. Artificial channels had been constructed to stabilize river flow paths before the occurrence of landslides in the lower part of the channel. Moreover, the side banks and riverbed of the channel were covered by concrete. This artificial channel is located 550 m from the sea and has a depth and width ranging from 2.5 to 3 m and from 10 to 20 m, respectively. The distance and flow line from the top of the landslide to the national road and the sea are 1,060 and 1,160 m, respectively.
4. Field Survey

4.1. Methods

We conducted four surveys to clarify the material properties and runoff and depositional processes of landslide-induced debris flows. We measured (a) the porosity of the soil and weathered bedrock on the hillslope and (b) the grain size distribution of the deposited materials to survey material characteristics. We then surveyed (c) the spatial patterns of erosion and deposition using aerial photographs and LiDAR data and (d) the runoff behavior using video images to clarify runoff and depositional processes.

4.1.1. Porosity Measurement for Soil and Weathered Bedrock in the Source Area

We used 20 intact 90- to 190-cm³ core samples from one borehole 50 m from the landslide scar (Figure 2c). The core samples were taken through soils and rocks. The sampling depths ranged from 1.7 to 36.4 m. We took a single sample at each depth. The numbers of the samples of volcanic soils, sands and gravels, welded tuff, and tuffaceous sedimentary rocks were eight, six, five, and one, respectively. We measured the saturated weight, underwater weight, and dry weight of both the soil and weathered bedrock to quantify their porosity.

4.1.2. Grain Size Analysis for Deposited Sediment

We conducted the following four surveys to clarify the grain size distribution from large boulders to fine sediments: areal sampling using an aerial photograph, grid sampling on the survey line, areal sampling using a ground-based photograph, and volumetric sampling.

First, the distribution of boulders larger than 100 cm was sampled using an aerial photograph taken on 30 July 2015. The interpreted area covered the part where most of the sediments were deposited (Figure 2c). We evaluated the percentile particle size of all boulders over 1 m on the basis of the occupied area for each boulder size (see Figure B1).

For boulders 15–100 cm in size, we set a 100-m survey line at a site around 610 m downstream from the landslide (Site A in Figure 2c) on 17 September 2015. The survey line was set on the deposited sediments along the flow direction. The grain diameters were measured at 1-m intervals along this line. On the basis of the number of grid points where each particle size was located, the particle size distribution of the 15–100-cm boulders was evaluated according to the number frequency of the particle per size class (see Figure B1).

Moreover, we took nine ground-based photographs and three sediment samples of deposited material on 17 September 2015, at Sites A and B, respectively (Figure 2c). Each photograph covered a 1 × 1-m square. The particle size distribution of the 5.3- to 15-cm deposits was measured on the basis of the area occupied by each size of sediment in the ground photograph. Next, we conducted volumetric sampling and physical sieving tests for sediments finer than 5.3 cm to quantify the frequency by weight of each size class.

Because previous studies had three sampling methods (i.e., grid, areal, and volumetric sampling) and two analyzed categories (i.e., by number and by weight), various conversion methods were proposed to obtain size distributions with the same sample and analysis categories (e.g., Bunte & Abt, 2001). Kellerhals and Bray (1971) used a model deposit comprising a mixture of three cube sizes packed without voids (i.e., a voidless cube model) to propose conversion methods between the various combinations of sampling methods and sample analyses. They concluded that the grid-by-number procedure is the only surface-oriented procedure directly comparable (equivalent) to customary bulk sieve analysis. However, several studies criticized that the voidless cube model over-simplified actual field conditions and proposed different conversion and photo-sieving methods (e.g., Ibbeken & Schleyer, 1986; Marion & Fraccarollo, 1997).

In this study, we used grid sampling for boulders 15–100 cm in size, volumetric sampling for those less than 5.3 cm in size, and aerial sampling for those over 100 cm and 5.3–15 cm in size. We used the measured areal occupancy ratio of each particle size class as its frequency by weight. Here, we considered that if the grid size is very small, the areal occupancy ratio of a given size class should be equivalent to the ratio of the occupied grid point of the class. This indicated that the ratio of the occupied area was directly comparable with that of the bulk sieve analysis. Therefore, we did not conduct any conversion to connect the data measured using different methods.
We measured the ratio of the occupied area of boulders over 100 cm to the entire interpreted area in the aerial photograph to quantify the frequency by weight of these boulders in the total deposited sediments (see Figure B1). Then, along the survey line, the occupancy ratio of the grid points with 15-cm sediments to the grid points with sediments less than 100 cm in size was quantified to evaluate the frequency by weight of particles smaller than 15 cm. Finally, the ratio of the occupied area of sediments smaller than 5.3 cm to the occupied area of sediments smaller than 15 cm was quantified using ground-based photographs to connect the particle size distribution above 5.3 cm with the particle size distribution below 5.3 cm (see Figure B1).

### 4.1.3. Topographic Analysis

We used aerial photographs and LiDAR data measured on 8 and 30 July, 2015. The spatial resolution of the LiDAR data is 1 m. No other landslides occurred during the interval of these LiDAR data, except for the two reference landslides. We can therefore evaluate the spatial patterns of net erosion and deposition depth due to debris flows induced by the two reference landslides and estimates the debris flow runout area. However, after the two reference landslides, the local government removed the deposited sediments in the artificial channel. Therefore, we needed to use video images to survey the deposition conditions in and around the artificial channel. We interpreted the images taken by Video Cameras 3 and 4 (see Figure 2c). We found that the artificial channel was almost filled by deposited sediments and that small parts of the debris flow overflowed the channel. Consequently, we assumed that the ground surface elevation after the debris flow runout in the artificial channel was the same as that outside the artificial channel at the section between the artificial channel’s upper end and the national road.

Because the landslide hillslope was very steep and the plan projected landslide area was limited, we considered the significant uncertainty of the calculated landslide volume from the LiDAR datasets. We evaluated the deposited volume at the section between the landslide toe and the national road using LiDAR datasets and quantified the discharge volume at the national road site using video images (see Section 4.1.4). Then, we integrated these two volumes to obtain the landslide volume. Using these data, we calculated the longitudinal change of the discharged volume due to debris flow including sediment and water at 20-m intervals.

### 4.1.4. Video Image Interpretation

We interpreted the video images (Video Cameras 1 through 4) taken by the central and local governments. The available view angles are shown in Figure 3. The video images were provided by Kyushu Regional Bureau, Ministry of Land, Infrastructure, Transport and Tourism (MLIT) and are available at the site of MLIT (http://www.qsr.mlit.go.jp/n-topics/h27/150813/tarumizu.html). The time-series of some images of the first debris flow captured by Video Cameras 1 through 4 are shown in Appendix A. Video Cameras 1 and 2 were installed at the same point and were used to monitor the steep hillslope where the landslide occurred. Video Camera 1 zoomed on the landslide, whereas Video Camera 2 monitored a wider area (Figures 3a and 3b). A clear V-shaped outcrop is located 60 m downstream from the landslide toe. Thus, we interpreted the start, peak, and end times of the debris flows induced by the two reference landslides at this outcrop using the video images from Video Camera 1.

We interpreted the debris flow propagation and stopping processes at Sites C and D using the video images from Video Camera 3 (Figure 3c). During the first debris flow, Video Camera 3 monitored the section between 750 and 780 m downstream from the landslide (Site D in Figure 2c). During the second debris flow, Video Camera 3 focused on the section between 660 and 680 m downstream (Site C in Figure 2c) because the local government changed its view angle. Video Camera 4 monitored the propagation and stopping processes in the region between 1,020 and 1,060 m downstream from the landslide (Site E in Figure 2c) during both the first and second debris flows (Figure 3d). We interpreted the front arrival time and flow end time at each site. The debris flow velocities were measured by interpreting the trajectory of boulders in the debris flows. We quantified the passing-through times of boulders at 60, 20, and 15 m distances in Sites C, D, and E, respectively. The elevation of the debris flow surface at Site E was also measured using the difference between the height of the artificial channel and that of the debris flow surface.

We interpreted the temporal change in the debris flow composition at Sites C, D, and E using Video Cameras 3 and 4. We set four transversal lines at 5- to 7-m intervals and provided seven observation points on each transversal line within the width of the debris flow (Figure 3c). We thus set 28 observation points for each site, allowing us to interpret whether the entire flow surface of each observation point was dominated by laminar or turbulent surges every 5–15 s during the debris flow runout. When boulders moved at the same flow velocity
as the mudflow or the flow surface appeared smooth, the flow was interpreted as a laminar surge (Figures 4a and 4b). The flow was interpreted as a turbulent surge if gravels moved separately from the mudflow, when boulders flowed down while floating and sinking on the water's surface, or when the flow caused splash and had an irregular surface (Figure 4a). Because the interpretation is based on the water surface condition, flows where the bottom is turbulent and the surface is smooth, such as plug flows, may be judged as laminar flows.

Figure 3. View angles of Video Cameras (a) 1, (b) 2, (c) 3, and (d) 4. The observation points for the debris flow composition at Site D are shown in panel (c).

Figure 4. Examples of interpreted turbulent and laminar surges at (a) Site C and D and (b) Site E.
4.2. Field Survey Results

4.2.1. Porosity of the Soil and Weathered Bedrock in the Source Area

Volcanic soils, sand and gravel layers, welded tuff, and tuffaceous sedimentary rocks respectively occur at depths of 0–10.5, 10.5–22.3, 22.3–35, and 35–36.5 m from the ground surface at the edge of the plateau close to the landslide scar. The porosity of the soil and weathered bedrock varied from 0.20 to 0.42 in the core samples, with average porosities for each lithology of 0.38, 0.28, 0.32, and 0.28, respectively (Figure 5). Except for the sand and gravel layers, there were no vertical porosity distributions in each layer. The porosity of the sand and gravel layers decreased with increasing depths. A visual observation of the boring core showed that the sand and gravel layers were composed of past debris deposits. Further, the observation indicated that whereas sand and gravel dominate in the upper layer, silt and clay dominate in the lower part. The mean porosity of the soil and bedrock averaged from the ground surface to the bottom of the landslide slip surface (i.e., volcanic soils, sand and gravel layers, and welded tuff) was around 0.33.

4.2.2. Grain Size Distribution of Deposited Sediment

We evaluated the entire grain size distribution of the debris-flow-deposited sediments in the study area as described in Section 4.1.2. Using the aerial photograph survey, we found 105 boulders larger than 1 m and quantified the ratio of the area covered by these boulders as 1%. The maximum diameter in the deposited area of the debris flow was around 4 m (Figure 6a). We found that the ratio of the boulders between 15 cm and 1 m in size to those smaller than 1 m was approximately 82% along the line survey at Site A. The grain size distribution between the 15-cm and 4-m boulders, combining the aerial photograph and the line survey, is shown in Figure 6a. The ground-based photograph interpretation showed that the ratios of the sediments smaller than 53 mm to those smaller than 15 cm did not greatly vary, ranging from 70% to 100% at Site B, whereas the ratios at Site A varied in the range of 35%–100% (Figure 6b). The mean grain size distribution at Site B was finer than that at Site A. The sieving test showed a limited difference between the grain size distributions for sieve-tested sediments smaller than 53 mm at Sites A and B. Around half of the sediments smaller than 53 mm were also smaller than 1 mm (Figure 6c). The entire grain size distribution of the deposited sediments obtained by combining the grain size distribution from the aerial photographs and the mean grain size distributions from the ground-based photography and the sieve test is shown in Figure 6d. This showed that grain size distribution ranged from around 0.1 mm to 4 m and that around 50% of the sediments had grain diameters approximately 3 mm (Figure 6d). Around 25% of the sediments were larger than 100 mm, and 35% were smaller than 1 mm.

4.2.3. Topographic Change

The spatial patterns of net erosion and deposition depth due to the debris flow are shown in Figure 7, where deposition in the artificial channel was not noted because of sediment removal.

A part of the sediments flowed into the sea, indicating that the debris flow reached the coast. The debris flow travel distance (distance from the landslide top to the lower ends of the deposits; L) and height (height from the landslide top to the lower ends of the deposits; H) were at least 1,160 and 200 m, respectively. Consequently, the value of H/L might be 0.17 or less.

Topographic analysis indicates that the debris flow runout area can be divided into three sections in terms of erosion/deposition patterns: the upper reach (i.e., the section between the landslide and 310 m downstream), the middle reach (i.e., the section between 310 and 610 m downstream), and the lower reach (i.e., the section between 610 and 1,060 m downstream; Figures 7 and 8). Appreciable erosion had occurred in the upper reach, whereas...
deposition dominated in the middle and lower reaches. Particularly, apparent deposition was observed in the middle reach. Conversely, deposition was minor and generally limited to the artificial channel in the lower reach.

In the upper reach, the runout area width interpreted from the aerial photographs ranged from approximately 30 to 100 m (Figure 8a). The maximum erosion depth was 2.6 m, and the maximum erosion volume along the flow direction was greatest (106 m$^3$/m) at 260 m downstream from the landslide (Figure 8b). The longitudinal gradient in the erosion section ranged from 7° to 34°, and the total erosion volume in the upper reach was around 5,800 m$^3$, including voids.

In the middle reach, two main areas of deposition were identified. First, around 19,000 m$^3$ of sediment, including voids, were deposited between 310 and 390 m, and the width of the first main deposition ranged from 80 to 110 m (Figure 8a). The maximum deposition depth was 3.9 m around 330 m downstream from the landslide (Figure 8b). Because this section was previously a quarry, it was relatively flat, with a slope gradient ranging from 1° to 2°, before the landslide occurrence. The second main deposition occurred at 430–610 m downstream from the landslide, and the width ranged from 70 to 120 m (Figure 8a). The longitudinal gradient was 2°–12° before the landslide. The lower end of this deposition occurred at almost the same position as that in the upper end of the artificial channel. The volume of the sediment deposits in this section was about 29,000 m$^3$, including voids, and the maximum deposition depth was 6.5 m at 510 m downstream from the landslide.

In the lower reach, no accurate data were available for the channel. Still, we confirmed that the net deposition is very small outside the artificial channel from images taken by Video Cameras 3 and 4. Moreover, there was no eroded area outside the artificial channel. The total volume of deposition in the lower reach was estimated at around 17,000 m$^3$. The deposition depth in the section ranged from 2.5 to 3 m; that is, the deposition depth was homogeneous compared with that in the middle reach.

Figure 6. Grain size distribution from the (a) aerial photograph interpretation and line survey, (b) ground-based photographs, and (c) sieve tests and (d) the combination of all the grain size survey results.
According to the spatial patterns of the net erosion and deposition depth in the debris flow runout area, together with the video interpretation, the total amount of sediment outflow, including water from the landslide scar, was estimated at approximately 58,700 m$^3$ (Figure 8c). Then, the total amount of sediment outflow increased because of erosion, reaching around 66,000 m$^3$ at the boundary of the upper and middle reaches (Figure 8c). Thereafter, the magnitude of sediment passage was reduced by deposition. Thus, the total amount of outflow at the boundary of the middle and lower reaches was 17,000 m$^3$. Deposition occurred in the lower reach, and the total amount of sediment outflow, including water from the lower reach, was calculated as 2,600 m$^3$.

4.2.4. Video Image Interpretation

4.2.4.1. Landslide and Debris Flow in the Upper Reach

According to the interpretation of the images from Video Camera 1, the first landslide occurred at 12:42:52 LT (Figure 9a), and the second occurred at 12:48:32 LT. These landslides became debris flows immediately after the landslide began. The first and second surges reached the V-shaped outcrop 60 m downstream from the landslide respectively 4 and 3 s after the landslide occurred (Figure 9b). The peak discharges of the first and second surges were observed from 5 to 9 s (Figure 9c) and 7 s after each landslide at the site. The durations of the first and second surges were 43 and 91 s, respectively. During the first surge, the discharge rate drastically reduced for 20 s.
4.2.4.2. Debris Flow in the Upper Part of the Lower Reach

The interpretation of the video images showed that the travel times of the first surge from the bottom of the landslide scar to Site D and the second surge to Site C were around 61 and 66 s, respectively (Figure 10a). We also found that the durations of the first surge at Site D and the second surge at Site C were 118 and 60 s, respectively. Turbulent flow was observed only at the front of the surges, and the durations of the turbulent flow in the first and second surges were 7 and 10 s, respectively. According to Video Cameras 3 and 4, the maximum surface velocity occurred at the front of both surges (Figure 10b). The velocity of the first and second surge fronts were 7.5 and 6.7 m/s at Sites D and C, respectively.

Thereafter, the ratio of the turbulent portion dramatically reduced, after which both surges became completely laminar flows in 3 and 10 s after the arrival of the first and second surges (Figure 10a). Then, a complete laminar flow persisted for up to 50 and 25 s, respectively. The surface velocity of the complete laminar flow in the first and second surges ranged from 2.4 to 6.7 m/s and 1.3 to 4.0 m/s, respectively (Figure 10b).
A portion of the flow then came to rest, and the stoppage area increased in size. Finally, the entirety of each of the surges stopped 118 and 75 s after first and second arrivals, respectively.

4.2.4.3. Debris Flow in the Lower Part of the Lower Reach

The travel times of the first and second surges from the bottom of the landslide scar to Site E were approximately 124 and 118 s, respectively (Figure 10a). At Site E, turbulent flow was observed at the front of the first surge. The duration of the entire turbulent flow in the first surge was long (30 s) compared with the corresponding value at Site D. The maximum velocity of the first surge was 5 m/s at the front of the surge (Figure 10b). Thirty seconds after the front's arrival, the surge changed to comprise mixed turbulent, laminar, and stationary portions. The turbulent portion decreased, and its surface velocity gradually declined with time. Finally, the entire surge stopped 60 s after surge arrival. Some of the first surges were deposited at Site E. As a result, the surface elevation of Site E gradually increased by approximately 2 m (Figure 10c). During the second surge, turbulent flow was not observed (Figure 10a). The video images showed that when the front of the second surge reached just upstream of Site E, the sediment deposited in the artificial channels because of the first surge began to be removed as laminar flows. Thus, an entirely laminar flow was observed at the surge front, lasting for up to 27 s after the front's arrival. The maximum velocity of the second surge was 2 m/s at its front (Figure 10b). A stationary portion appeared and increased with time. The surface velocity decreased, and the entire surge stopped 41 s after arrival. The surface elevation dramatically increased by approximately 1 m just after the front of the second surge.
The first debris flow surge transitioned from laminar flow into turbulent flow in its lower reach (Figure 10). The deposited sediments became finer in the lower site than that in the upper site in the lower reach (see Figure 6b). These indicate that although the debris flow's water content should be only around 30% just after the landslides occurred at the survey site (see Figure 5) and although the upper and middle reaches incorporated a substantial
amount of large boulders to the debris flow, a fluid phase nonetheless dominated the debris flow downstream and delivered a large amount of fine sediments to the lower reach.

The flow behaviors in the upper part of the lower reach were similar in the first and second surges (Figure 10). Moreover, no obvious differences were observed between the travel times from the upper and lower parts in the lower reach between the first and second surges. This suggests that the flow behavior in the second surge is similar to that in the first surge, albeit masked due to the removal of the deposited sediments.

5. Numerical Simulation

5.1. Numerical Simulation Methods

5.1.1. Simulation Program

The video images indicated that the landslide transformed into a debris flow immediately after its occurrence. We conducted alternative physically based numerical simulations using Kanako-LS (Uchida et al., 2013), a program developed for describing large-scale stony debris flows. These were done to investigate the roles of phase-shift and depositional processes on the runoff propagation of the landslide-induced debris flow. Kanako-LS is based on shallow water equations for granular flows and a 1D model. The physical, mathematical, and computational bases of Kanako-LS are shown in detail in Appendix C and described by Uchida et al. (2013, 2021). Kanako-LS is a single-phase model that evaluates erosion and deposition rates and riverbed shear stress based on the stony debris flow theories proposed by Takahashi and colleagues (e.g., Takahashi, 2009; Takahashi & Nakagawa, 1991). They proposed that bed shear stress might be dominant in terms of flow resistance. Further, they divided flow conditions into three classifications in terms of sediment concentration: stony debris flow, immature debris flow, and turbulent water flow. They also proposed equations to evaluate each bed shear stress (Equations C13–C15). We included depositional processes of phase-shifted sediment into the numerical model developed by Uchida et al. (2013) using Equations 10 and 11.

Recently, two-phase debris flow models have been widely applied to describe the complex dynamics of two-phase debris flows by considering solid momentum, fluid momentum, and interfacial momentum transfer (e.g., Pudasaini, 2012; Tai et al., 2019). These two-phase debris flow models included various parameters, such as the drag coefficient, the viscosity of the interstitial fluid, and the fluid friction coefficient. However, setting values for these parameters is difficult for actual debris flows. Thus, previous studies conducted comprehensive sensitivity analyses of such parameters (Tai et al., 2019). In contrast, the single-phase debris flow models have also been verified by experiments to reproduce steady-state debris flow propagation well under various conditions (e.g., Sakai et al., 2019; Takahashi, 2009). Further, they have reproduced actual past debris flows well (e.g., Takahashi et al., 2001; Uchida et al., 2021). Moreover, Uchida et al. (2021) successfully clarified the role of phase-shifted sediments on debris flows by numerically simulating an actual debris flow using a single-phase model. Thus, we thought that single-phase models would probably more easily determine the role of phase-shifted sedimentary deposition processes in the propagation of debris flows.

Phase-shifted sediments are assumed to be deposited as skeleton-forming sediments in the original Kanako-LS, as are non-phase-shifted sediments. In this study, we included the proposed hypothesis for phase-shifted sediment depositional processes (see Section 2) into Kanako-LS.

5.1.2. Parameter Setting

The simulation was applied for the section from the bottom of the landslide scar to the coast. Using aerial photographs and LiDAR data, we set the width and the longitudinal ground surface profile of the debris flow runout area at 20-m intervals (Figures 8a and B1). We used the valley bed width before the debris flow estimated from the cross-section topographic profile and the debris flow trace width interpreted from the aerial photographs to set the path width of the debris flow. For the simulation, the debris flow width was taken as the average of these two widths (Figure D1 in Appendix D). We assumed that, in the upper and middle reaches, the initial depth of the movable layer at the ground surface was equal to the maximum observed erosion depth, which ranged from 1 to 3.5 m. In the lower reach, the initial depth of the movable layer was set to 0 m, and we modeled the artificial channel. This is because field data suggest that the amount of overflow from the artificial channel was small. The values of other primary simulation parameters were set as shown in Table 1. We assumed that the landslide sediments were fully saturated with water because many water seepage features were observed in the landslide.
scar. Although a reduction in porosity upon failure of the initial landslide mass and induced excess pore water pressure is a possibility, we simply assumed here that the porosity of the debris flow was similar to that of the initial landslide mass. Thus, the water content of the debris flow was set as the average porosity of both the soil and bedrock (0.33). We set the concentration of the skeleton-forming sediments in the deposits \((C_{ds})\) as 0.67, similar to that of the landslide sediment. Moreover, we did not consider spatial and temporal variations in \(C_{ds}\). This indicates that the concentration of the skeleton-forming sediments in the deposits is longitudinally constant in the simulation section regardless of deposition and erosion. The coefficient of the erosion rate was set to the general value of 0.0007 (Takahashi, 2007). Although the coefficient of the deposition rate affects the simulation result, it was set to 0.05, following Takahashi (2007), Takahashi and Nakagawa (1991), Nishiguchi et al. (2014), Nakatani et al. (2016), Suzuki et al. (2019), and Uchida et al. (2021). The Manning’s coefficient in mountain rivers vary greatly from basin to basin (Asano et al., 2018). It is often empirically set at between 0.03 and 0.06 \(m/s^{1/3}\) in the simulation of debris flows (e.g., Nakatani et al., 2016; Takahashi et al., 2001; Uchida et al., 2013). In this study, we used an intermediate value (0.045). The time step for the calculation was set to 0.005 s as the standard value. We confirmed that even if the time step was set to 0.001 s for a few cases, the difference in the height of the riverbed change was only a few percent, and that there was no large difference. The minimum depth of debris flow was set to 0.01 m. If the water depth is less than 0.01 m, we assumed no flow occurred.

We used data for the debris flow duration, peak time, and start time of the gradual decrease of the hydrograph at the bottom of the landslide scar to set the input hydrograph (see Section 4.2.4.1). Thereafter, we assumed that the debris flow hydrographs linearly increased and decreased and that the starting discharge is 0.1 times that of the peak discharge, as shown in Figure 11. The peak velocities of the first and second surges at Site A were about 8 and 7 m/s, respectively. These suggest that there were no apparent differences in the peak discharge. We therefore set the same peak discharge for both surges. The inflow position for the input hydrograph was the bottom of the landslide scar (Figure D1).

We set four patterns of the ratios of the phase-shifted sediments in the debris flow to the total sediment (hereafter, the phase-shift ratio) at the upper end of the simulation section: 0%, 20%, 30%, and 40%. This was done to investigate the roles of the phase-shifted sediments included in the debris flow. We did not consider the different grain size distributions between the landslide mass and the eroded materials during the debris flow runout. However, we assumed that sediments smaller or larger than a given boundary diameter behaved as phase-shifted or non-phase-shifted sediments, respectively. Thus, according to the grain size distribution of deposited sediments (see Figure 6), the sediments...
Table 2
Simulation Cases

| Case | k  | Phase-shift ratio | \( C_{00} \) | \( C_{01} \) | \( d_{0} \) (m) | \( d_{4} \) (mm) |
|------|----|------------------|-------------|-------------|---------------|---------------|
| 1a   | 0  | 0                | 0.67        | 0.000       | 0.094         | 0.000         |
| 1b   | 0.2| 0.536            | 0.134       | 0.289       | 0.117         | 0.115         |
| 1c   | 0.3| 0.469            | 0.201       | 0.379       | 0.134         | 0.197         |
| 1d   | 0.4| 0.402            | 0.268       | 0.448       | 0.156         | 0.620         |
| 2a   | 0.5| 0                | 0.67        | 0.000       | 0.094         | 0.000         |
| 2b   | 0.2| 0.536            | 0.134       | 0.289       | 0.117         | 0.115         |
| 2c   | 0.3| 0.469            | 0.201       | 0.379       | 0.134         | 0.197         |
| 2d   | 0.4| 0.402            | 0.268       | 0.448       | 0.156         | 0.620         |
| 3a   | 0.75| 0               | 0.67        | 0.000       | 0.094         | 0.000         |
| 3b   | 0.2| 0.536            | 0.134       | 0.289       | 0.117         | 0.115         |
| 3c   | 0.3| 0.469            | 0.201       | 0.379       | 0.134         | 0.197         |
| 3d   | 0.4| 0.402            | 0.268       | 0.448       | 0.156         | 0.620         |
| 4a   | 0.875| 0              | 0.67        | 0.000       | 0.094         | 0.000         |
| 4b   | 0.2| 0.536            | 0.134       | 0.289       | 0.117         | 0.115         |
| 4c   | 0.3| 0.469            | 0.201       | 0.379       | 0.134         | 0.197         |
| 4d   | 0.4| 0.402            | 0.268       | 0.448       | 0.156         | 0.620         |
| 5a   | 1  | 0                | 0.67        | 0.000       | 0.094         | 0.000         |
| 5b   | 0.2| 0.536            | 0.134       | 0.289       | 0.117         | 0.115         |
| 5c   | 0.3| 0.469            | 0.201       | 0.379       | 0.134         | 0.197         |
| 5d   | 0.4| 0.402            | 0.268       | 0.448       | 0.156         | 0.620         |

Note. \( C_{00} \): initial coarse sediment concentration in the debris flow; \( C_{01} \): initial fine sediment concentration in the debris flow; \( d_{0} \): initial fine sediment diameter in the interstitial water; \( d_{4} \): representative coarse sediment diameter; \( d_{4} \): representative fine sediment diameter.

When the phase-shift ratio was 0% (Cases 1a through 5a), all simulated results were the same. This suggests that around 58,000 m³ of sediment, including water, was deposited and stopped between the landslide and 120 m downstream of the landslide in the upper reach (light blue lines in Figures 12a–12e). When the phase-shift ratio was 0.2 (Cases 1b through 5b), deposition did not seem to have occurred in the upper reach (orange lines in Figures 12a–12e). Apparent deposition occurred in the upper part of the middle reach, between 310 and 370 m downstream from the landslide. The total volume of the deposited sediments ranged from 26,000 to 37,000 m³ (orange lines in Figures 12a–12e and 13a–13e). The total amount of outflow from the section above was almost 0 in Case 1b (Figure 13a) but increased with increasing values of \( k \) (Cases 2b through 5b). At the phase-shift ratios of 0.3 and 0.4 (Cases 1c through 5c and 1d through 5d), erosion was dominant in the upper reach, and apparent deposition occurred twice within the middle reach (red and blue lines in Figures 12a–12e). The deposition volume between around 310 and 370 m downstream from the landslide decreased compared with those where the phase-shift ratio was 0.2. The total amount of outflow from the middle reach increased with increasing phase-shift ratios and increasing values of \( k \) (red and blue lines in Figures 13a–13e).
5.2.1.2. Lower Reach

When $k$ was set to 0 (Case 1), no significant erosion and deposition due to debris flows were observed in the lower reach (Figures 12a and 12b). This is because the total amounts of outflow at the lower end of the middle

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Figure 12. (a–e) Simulated and observed riverbed change volumes per unit distance due to the debris flow. The left figures show the range from the landslide scar to the sea (1,200 m downstream of the scar); the right figures show an extended range from 500 to 1,100 m downstream of the landslide scar. The letters in the legend refer to each calculation case shown in Table 2.
reach were less than 2,400 m³ regardless of the phase-shift ratio (Figures 13a and 13b). In contrast, when $k$ was set to 0.5, 0.75, 0.875, and 1.0, the deposition volume increased with increasing phase-shift ratios when $k$ was kept constant (Figure 12b through Figure 12e). Moreover, the deposition volume was largest when $k$ was set to 0.875, compared with those in Cases where $k$ was set to 0.5, 0.75, or 1.0. When $k = 1.0$, much sediment reached the coast without deposition in the lower reach (Figure 13e). In contrast, when $k$ was 0.5, most of the outflow in the lower reach was water (Figure 13c through Figure 13d).

5.2.1.3. Composition of Deposits

The volumetric ratios of the coarse sediments, fine sediments, and water in the deposited sediments in the main deposit area obtained from the simulation results were calculated. The main deposit area in Case a ranged from the landslide scar to 120 m downstream (Figure 12). The main deposit area in Cases b–d ranged from 320 to 360 m and from 460 to 1,160 m downstream of the landslide except for Case 5d. That in Case 5d ranged from 320 to 360 m and from 540 to 1,160 m downstream of the landslide (Figure 12).

When the phase-shift ratio was 0 (Cases 1a through 5a), the deposits comprised only coarse sediments. The ratio of the sediments in the deposits was 0.67, and that of water was 0.33 (Figure 14a). When the phase-shift ratio was set to 0.2–0.4, as $k$ increased, the ratio of the total sediment (including both fine and coarse sediments) increased, and the ratio of water decreased. When $k$ was zero and the phase-shift ratio ranged from 0.2 to 0.4 (Cases 1b through 1d), the ratio of the total sediments in the final deposits was 0.67. The ratios of the coarse and fine sediments for each phase-shift ratio were the same as those in the landslide mass and initial riverbed for the same phase-shift ratio (Figures 16b–16d). When $k$ was 1 and the phase-shift ratio ranged from 0.2 to 0.4 (Cases 5b through 5d), the ratio of the total sediment was 0.77–0.82, and the ratio of water was 0.18–0.23 in the final deposits. This is because of the increase in the void-space-stored sediments. The ratios of the void-space-stored sediments for the Cases when the phase-shift ratio was set to 0.2, 0.3, and 0.4 were 0.10, 0.13, and 0.15, respectively (Figures 14c and 14d).

The simulated results showing that the debris flow travel distance was dependent on the depositional processes of fine sediments may be explained as follows. When $k$ increases, the concentration of sediments stored in the void spaces of the final deposit increases; that is, the total sediment concentration in the final deposit increases. Thus, if $k$ is large and close to 1, the sediment concentration in the debris flow decreases because of debris flow deposition, and the debris flow becomes a fluid phase-dominated flow. This fluid phase-dominated flow was present even in the lower reach. In contrast, when $k$ is small and close to zero, sediment concentrations in the debris flow should remain high because the total sediment concentration in the final deposit was the same as those in the landslide mass and initial riverbed. Thus, the debris flow did not behave as a “flow” in the lower reach, where longitudinal gradients are low.

5.2.2. Comparison of the Simulated and Observed Results

The relative errors, $E$, within the upper and the middle reaches were less than 0.2 in Cases 2c, 3c, 4b, and 5b (Figure 15a). In these Cases, apparent erosion occurred between the landslide and 310 m downstream, and apparent deposition occurred between 310 and 370 m and between 430 and 610 m.
Figure 14. The ratios of the coarse sediment, fine sediment, and water in the initial bed and in the debris deposits for (a) Cases 1a through 1d, (b) Cases 1b through 5b, (c) Cases 1c through 5c, and (d) Cases 1d through 5d. Below the dotted line, the constituent materials of the skeleton-forming sediments and their ratios are shown. Above the dotted line, the constituent materials of the voids of the riverbed and their ratios are shown. The letters in the legend refer to each calculation case shown in Table 2.
downstream of the landslide (Figures 12b–12e and 13b–13e). These simulated erosion and deposition patterns agreed well with those in the observations.

Cases 2b, 2c, 2d, 3b, and 4b exhibited small $E$ values for the lower reach (less than 0.6; Figure 15b). In Cases 2b, 2c, and 3b, almost no deposition occurred in the artificial channel. Meanwhile, in Cases 2d and 4b, maximum deposition up to about 1 m occurred throughout the section (Figures 12b–12d and 13b–13d). Although the deposition depth of the simulation was slightly less than the observational depth, the simulated deposition patterns were almost the same as the observed deposition patterns.

In Case 4b, which accurately reproduced the observed results for both the upper and middle reaches and the lower reach, the ratio of the total amount of the phase-shifted sediment outflow to that of the overall sediment outflow ranged from 20% to 40% in the section between the landslide and 500 m downstream (Figure 16). The ratio increased in the lower reach and was 99% at Site E, delineating that the fluid phase was dominant. This is also consistent with results from the video image interpretation of the first surge at Site E. Around 56% of the deposited fine sediments were skeleton-forming sediments, and the rest were void-space-stored sediments (Figure 14b).

These results suggest that during debris flow deposition, some phase-shifted sediments may have been stored in the interstitial spaces of the riverbed, resulting in a fluid phase-dominated flow. In the upper and middle reaches, the $E$ values were 0.6–0.8, 0–0.6, 0–0.4, and 0.2–0.8 when the phase-shift ratios were 0, 0.2, 0.3, and 0.4, respectively. We found that $E$ varies with the phase-shift ratio. However, differences in $E$ due to differences in $k$ were relatively small (Figure 15a). In contrast, in the lower reach, the maximum values of the difference in $E$ due to differences in $k$ were 2.7 and 4.5 when the phase-shift ratios were 0.3 and 0.4, respectively. We also noted a significant variation in $E$ (Figure 15b). Therefore, the reproducibility of the upper and the middle reaches was more affected by the phase-shift ratio than it was by $k$. In contrast, the reproducibility of the lower reach was significantly affected by $k$ in addition to the phase-shift ratio.

6. Discussion

6.1. Roles of Depositional Process of Fine Sediment in Travel Distance

The transformation of landslides into long runout debris flows was explained using the concept of the phase-shift of sediments from the solid phase to the fluid phase. This results in effects such as increased buoyancy acting on a solid phase or reduced energy dissipation due to the decrease in both the collisions between particles and the drag between particles and the interstitial fluid (e.g., Egashira et al., 2016; Iverson & George, 2014; Nishiguchi et al., 2011; Tai et al., 2019; Uchida et al., 2013). Various numerical simulation methods introduced phase-shift effects of fine sediments in debris flows (e.g., Iverson et al., 2010; Iverson & George, 2014; Pudasaini, 2012;
Tai et al., 2019). Most previous numerical models that considered the depositional processes of phase-shifted sediments applied the relatively simple hypothesis that phase-shifted sediments become skeleton-forming sediments or void-space-stored sediments (i.e., \( k = 1 \) or \( 0 \) in this study). The influences of the depositional processes of phase-shifted sediments on the travel distance of debris flows have not yet been fully verified (e.g., Egashira et al., 2016; Takahashi et al., 2001). This study demonstrated that, when considering the effects of phase-shifted sediments on debris flows, patterns of erosion and deposition in the upper and middle reaches can be described. However, the travel distance and deposition amount of the entire debris flow, including the fluid phase-dominant flow in the lower reach, cannot be described in terms of the simple depositional processes of fine sediments (i.e., \( k = 1 \) or \( 0 \)).

This phenomenon could be well-described if most phase-shifted sediments in the debris flow are assumed to remain in a turbulent state on the riverbed during deposition (\( k = 0.875 \)). In other words, paying attention to both the phase-shift of fine sediments in debris flows and the depositional processes of phase-shifted sediments is necessary for estimating the travel distance or runout area of a landslide-induced debris flow.

We did not directly validate the depositional processes of the phase-shifted sediments in the field. Rather, we examined this hypothesis using numerical simulations. Hydraulic model experiments were conducted to investigate the effects of fine sediments on debris flow runout processes and deposit morphology (e.g., de Haas et al., 2015; Iverson et al., 2010; Zhou et al., 2019). Although the depositional processes of phase-shifted sediments have not been directly addressed to date, we found several lines of evidence for fine sediment depositional processes in debris flows from several previous model experiments. Major (2000), Frings et al. (2008), Iverson et al. (2011), and de Haas et al. (2015) argued that the rate of skeleton formation in beds during deposition depends on the grain size and the shape of the grain size distribution. This supports the hypothesis that relatively fine-grained phase-shifted sediments maintain their phase-shifted status and deposit as void-space-stored sediments, whereas other sediments form skeleton structures on the riverbed (i.e., \( k \) should be neither 0 nor 1) because the debris flow has a wide grain size distribution.

This is the case at Fukaminato River (Figure 6).

In addition, Iverson et al. (2010) experimented on the depositional processes of debris flows with high sediment concentrations (approximately 60%) using two types of sediments. In their experiment without fine sediment mixing, the debris flow was not fully liquefied. However, in the case of fine sediment mixing, although the debris flow was not fully liquefied during the flow-down, the debris flow nonetheless underwent liquefaction as deposition progressed and consequently traveled a longer distance than did the debris flow without intermixed fine sediments. Similar results were observed in the experiment by de Haas et al. (2015). However, when the clay content exceeded 0.22, the runout distance decreased because of the increased viscosity. These results are supported by the findings of this study: that the deposition of fine sediments contributes to increased concentrations of sediments deposited on the riverbed and enhanced fluid phase-dominated flow.

6.2. Ways Forward

However, we noted a discrepancy in the shape of the particle distribution between debris deposits of this study and fluvial deposits of Frings et al. (2008). Frings et al. (2008) reported that the percentage of void-space-stored sediments of the fluvial deposits was greater in bimodal sediments than in unimodal sediments. Moreover, they showed that the percentage of void-space-stored sediments became almost zero in unimodal sediments. In contrast, Figure 6 shows the unimodal deposited sediments in Fukaminato River. Although more detailed surveys of particle distribution are needed, we can conclude that the wide range of particle distribution (i.e., more than four orders of magnitude) of the debris deposits affected the relatively high percentage of void-space-stored sediments in the best-fit case (Case 4b). So, we think that the processes of the fluvial deposition cannot fully explain for the debris flow deposition processes, because the fluvial deposits are formed under very different hydraulic conditions, sediment concentrations, grain-size distributions, and grain sizes in comparison with debris flows, and that further research is needed on the determination mechanism of \( k \) of the debris deposits.

Figure 16. The simulated total amounts of non-phase-shifted and phase-shifted sediment outflows (black and red solid lines on the left axis) and the simulated ratio of the total amount of the phase-shifted sediment outflow to that of the overall sediment outflow (blue broken line on the right axis) in Case 4b.
In this study, we used relatively simple model, 1D and single-phase flow model. Recently, several 2D and two-phase flow model were developed and applied actual landslide induced debris flow (e.g., Tai et al., 2019). Moreover, in this study, we did not consider a variety of processes, such as the viscosity of intestinal fluid, the interfacial momentum transfer between solid and interstitial fluid, the momentum exchange because of erosion and deposition, the lateral and longitudinal particle segregation and so on, but recent models considered some of these processes (e.g., Iverson & George, 2014; Pudasaini, 2012; Tai et al., 2019). However, our simulation well described the propagation of actual landslide-induced debris flow, suggesting that these processes may be minor contribution of debris flow in Fukaminato River. For example, the narrow outflow channel in the lower reach of Fukamiato River constrained lateral flow and facilitated longitudinal flow, although several studies found lateral segregation particle size-segregation processes in debris flows (e.g., Johnson et al., 2012; de Haas et al., 2015). Therefore, we believe that comparing various models with different approximation methods of physical processes in debris flow, such as applying the single-phase flow model and the two-phase flow model to the same phenomenon, will promote the understandings of debris flow in the future. It is expected to contribute to disaster prediction and mitigation.

7. Conclusion

We presented a new data set obtained from a landslide-induced debris flow that occurred along Fukaminato River in July 2015. This included the initial state of the landslide, the transition from landslide to debris flow, and the runout and deposition of the debris flow.

Using these field data, we clarified that the landslide transformed into a debris flow immediately after its initiation. This debris flow induced erosion in the upper reach (longitudinal riverbed angle range of 7°–34°). Around 74% of the eroded sediment from the landslide and the upper reach was deposited in the middle reach (longitudinal riverbed angle range of 1°–19°). In the lower reach (longitudinal riverbed angle range of 1°–3°), the debris flow transitioned from laminar to turbulent. Because the studied landslide occurred during a rain-free period with no surface runoff observed, the water content of the debris flow should have been limited by the porosity of the landslide. Thus, the water content of the debris flow should have been approximately 30% just after the initiation of the landslides at the survey site. We inferred that the fine sediments behave as a part of the interstitial fluid in the debris flow and tested this inference using numerical simulations. The results of our observations and simulations indicated that a fluid phase-dominated flow propagated downstream and delivered a large amount of fine sediments in the interstitial fluid to the lower reach. Moreover, the fluid phase-dominated portion flowed over 1.2 km and caused extensive damage.

Our numerical simulations indicated that when we assumed that all fine sediments that behaved as fluid phase in the debris flow formed skeleton deposits with other coarse sediments, the debris flow did not transition to a fluid phase-dominated flow and did not travel long distance. However, when all fine sediments were assumed to be stored in the interstitial spaces of the riverbed with water, the debris flow became fluid phase-dominated in the lower reach. Still, the amount of deposition in the lower reach was underestimated. Thus, when we assumed that a portion of the fine sediments forms skeleton deposits with other coarse sediments and that others are stored in interstitial spaces, the deposition pattern in the lower reach could be described accurately. According to these results, we can conclude that fine sediment behavior in flow and deposition processes significantly impact erosion and deposition patterns and the travel distance of landslide-induced debris flows.

In this study, we set up a simple constant coefficient (\(k\)) to describe the depositional processes of fine sediments. Consequently, the spatial and temporal variations of this coefficient and its controlling factors have not been discussed. According to previous laboratory experiments (de Haas et al., 2015) and river sedimentology research (Frings et al., 2008), we can conclude that \(k\) is controlled by the shape of the particle size distribution. Further works to clarify these issues, such as analyses of other landslide-induced debris flows or hydraulic model experiments of debris flows, are desirable. To date, there are few studies that have observed actual debris flow deposits and confirmed the sediment that formed the skeleton and the sediment stored in the interstitial spaces of the riverbed. Future studies to observe the structure of the debris flow deposits in the field will improve knowledge of debris flow propagation and contribute to improving techniques for predicting potential damage areas.
Appendix A: Time-Series of Images Captured by Video Cameras 1–4

The time-series of some images of the first debris flow captured by Video Cameras 1 through 4 are shown in Figures A1, A2, A3, A4 and A5.

**Figure A1.** Time-series images every 2 s captured by Video Camera 1. The times (in seconds) displayed above each figure are the elapsed time after the first landslide.
Figure A2. Time-series images every 5 s captured by Video Camera 2. The times (in seconds) displayed above each figure are the elapsed time after the first landslide.
Figure A3. Time-series images every 5 s captured by Video Camera 3. The times (in seconds) displayed above each figure are the elapsed time after the first landslide.
Figure A4. Time-series images every 5 s captured by Video Camera 3 (continuation of Figure A3). The times (in seconds) displayed above each figure are the elapsed time after the first landslide.
Figure A5. Time-series images every 5 s captured by Video Camera 4. The times (in seconds) displayed above each figure are the elapsed time after the first landslide.

Appendix B: Clarifying the Entire Particle Distribution of the Deposited Sediments

The method used to clarify the entire particle distribution of the deposited sediments is shown by the dotted line in Figure B1.
Appendix C: Basic Governing Equations in Kanako-LS

The equation used to determine the change in the bed surface elevation can be written as follows:

$$\frac{\partial Z}{\partial t} + i = 0$$  \hspace{1cm} (C1)

where $i$ represents the erosion and deposition rate, and $z$ (m) is the height of the riverbed. Positive values of $i$ indicate erosion. In the modified Kanako-LS, we used three continuity equations for the entire flow, non-phase-shifted sediments, and phase-shifted sediments, as described in Equations 8, 11 and 12, respectively.

Then, the erosion and deposition rate were calculated by the entrainment equation proposed by Takahashi (Uchida et al., 2013), and erosion rate is given by Equation C2 when $C_\infty \geq C_c$. The deposition rate is given by Equation C3 when $C_\infty < C_c$.

$$i = \delta_e \frac{C_\infty - C_e}{C_{e_0} - C_\infty} \frac{q}{d_c}$$ \hspace{1cm} (C2)

$$i = \delta_d \frac{C_\infty - C_e}{C_{e_0} - \frac{q}{h}}$$ \hspace{1cm} (C3)

where $C_\infty$ is the equilibrium non-phase-shifted sediment concentration in the debris flow, $\delta_e$ and $\delta_d$ are the coefficients of the erosion and deposition rates, respectively, $q$ (m$^3$/s) is the debris flow discharge per unit width, and $d_c$ (m) is the representative diameter of the non-phase-shifted sediments.

Following Takahashi et al. (2001), we used a set of equilibrium sediment concentrations in the debris flow that can be represented as follows:

Stony type debris ($\tan \phi > \tan \theta_w \geq 0.138$),

$$C_\infty = \frac{\rho \tan \theta_w}{(\sigma - \rho)(\tan \phi - \tan \theta_w)}$$ \hspace{1cm} (C4)

where $\rho$ (kg/m$^3$) is the mass density of the sediment, $\sigma$ (kg/m$^3$) is the sediment density, $\phi$ (°) is the angle of the internal friction within the sediment, and $\theta_w$ is the inclination angle of the flow surface along the flow direction.

Immature type debris flow ($0.138 > \tan \theta_w > 0.03$),

$$C_\infty = 6.7 \left( \frac{\rho \tan \theta_w}{(\sigma - \rho)(\tan \phi - \tan \theta_w)} \right)^2$$ \hspace{1cm} (C5)
Turbulent water flow with bedload transport \((\tan \theta_w \leq 0.03)\),

\[
C_\infty = \left(\frac{1 + 5 \tan \theta_w}{\pi} \right) \left[ 1 - a_c^2 \frac{\tau_c}{\tau_c^*} \left( 1 - \frac{\tau_c}{\tau_c^*} \right)^{1/3} \right] \tag{C6}
\]

where \(\tau_c\) is dimensionless shear stress, and \(\tau_c^*\) is critical dimensionless shear stress, such that

\[
\tau_c^* = 0.04 \times 10^{1.72 \tan \theta_w} \tag{C7}
\]

\[
a_c^2 = \frac{2 \left( \frac{0.425 - \frac{\sigma \tan \theta_w}{\sigma - \rho}}{1 - \frac{\sigma \tan \theta_w}{\sigma - \rho}} \right)}{1 - \frac{\sigma \tan \theta_w}{\sigma - \rho}} \tag{C8}
\]

\[
\tau_s = \frac{\rho}{\sigma - \rho} \frac{h \tan \theta_w}{d} \tag{C9}
\]

where \(d\) (m) is the mean diameter of the movable sediments on the riverbed. When \(\tau_s \leq \tau_c^*\),

\[
C_\infty = 0 \tag{C10}
\]

Moreover, the mass density of the interstitial fluid can be written as

\[
\rho = \frac{\sigma C_l + \rho_w C_w}{C_l + C_w} \tag{C11}
\]

where \(\rho_w\) (kg/m\(^3\)) is the mass density of pure water.

Following Takahashi et al. (2001), the momentum equations for the flow phenomenon in the x-axis direction are as follows:

\[
\frac{\partial Q}{\partial t} + u \frac{\partial (uQ)}{\partial x} = -gA \sin \theta_w - \frac{\tau_x}{\rho} B \tag{C12}
\]

where \(g\) (m/s\(^2\)) is the gravitational acceleration, \(A\) (m\(^2\)) is the cross-sectional area of flow, \(\tau_x\) (N/m\(^2\)) is the riverbed shearing stress in the x-axis direction, \(\rho_i\) (kg/m\(^3\)) is the density of debris flow material, and \(B\) (m) is the debris flow width. In the equation, water surface gradient was used for describing the gravity component of the momentum equation and the equation has been applied to simulate a variety of debris flow and successfully described debris flow propagation (e.g., Nakatani et al., 2016; Takahashi et al., 2001; Uchida et al., 2013). The parameter \(\tau_x\) is described by Equations C13–C15 (Takahashi, 2007):

Stony debris \((C_c \geq 0.4 C_{ch})\),

\[
\frac{\tau_x}{\rho h} = \frac{u^2 d_c^2}{8 h^3 \left\{ C_c + (1 - C_c) \frac{\rho_i}{\sigma} \right\} \left\{ \frac{C_c}{C_l} \right\}^{1/3} - 1 \}^2 \tag{C13}
\]

Immature debris flow \((0.01 < C_c \leq 0.4 C_{ch})\),

\[
\frac{\tau_x}{\rho h} = \frac{1}{0.49} \frac{u^2 d_c^2}{h^3} \tag{C14}
\]

Turbulent water flow with bedload transport \((C_c \leq 0.01\text{ or } h/d_c \geq 30)\),

\[
\frac{\tau_x}{\rho h} = \frac{g n_m^2 u^2}{h^{4/3}} \tag{C15}
\]

where \(n_m\) is Manning’s coefficient (s/m\(^{1/3}\)).
Appendix D: Topography for the Numerical Simulation

The longitudinal profile, slope angle, and flow width of the simulated section are shown in Figure D1.

![Figure D1. Longitudinal profile, slope angle, and flow width of the simulated section.](image)

Data Availability Statement

The video images were provided by the Kyushu Regional Bureau, Ministry of Land, Infrastructure, Transport and Tourism (MLIT). The LiDAR data, aerial photographs, and porosity data of the soil and weathered bedrock on the hillslope in Fukaminato River were provided by the local (Kagoshima prefectural) governments. The porosity and geology data can be downloaded from https://data.mendeley.com/datasets/dp8wy6d45w/draft?a=8332e5fc-6e16-4cc4-8d54-15354cdbf644 (e-mail: nishiguchi@ctie.co.jp; name: yuki nishiguchi; password: data-j2021). The grain size data were provided by the National Institute for Land and Unfractured Management (NILIM) and are available for research purposes. The grain size data in Fukaminato River can be downloaded from https://data.mendeley.com/datasets/wcwxtdf6ng/draft?a=31ae2d01-1587-456c-972a-9511d4aae4d6. The topographic elevation data before and after the landslide and the debris flow width, net volume change per meter along the channel, and the total amount of outflow shown in Figure 8 can be downloaded from https://data.mendeley.com/datasets/r7bn9s9y6a/draft?a=56b11bdf-e01a-4c9d-bd32-c1654f7f45ca using the previous password. Although not used in this study, the 10-m digital elevation models before and after the landslide generated from contour lines created in 2015 and 2018, respectively, are available at the site of the Geospatial Information Authority of Japan (https://www.gsi.go.jp/kiban/index.html).

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