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SOME EFFECTS OF SLOPE MOVEMENTS ON RIVER CHANNELS

BY

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SYNOPSIS

The geomorphic response of a channel to slope movements depends on the rate of colluvium delivery from hillslopes relative to the rate of removal by fluvial processes. For slope movements ranging from extremely slow to slow (less than 1.5 m/mo). the balance between slope and fluvial influences can be considered in terms of the channel constriction ratio, defined as the ratio of slope movement to channel width, expressed as %/yr: Sites with faster constriction ratios may experience more frequent stream bank instability and sediment production. Rapid slope movements discharging large volumes of colluvium into channels can form dams. Darn volume, river discharge, and size distribution of dam matetial determine the potential for dam failure and release of water from impounded lakes. The implications for hazard prediction and mitigation are discussed.

INTRODUCTION

Slope movements of soil and rock (using the terminology of Varnes. 1978) are major environmental hazards of national proportions in Japan, and they also pose very significant hazards in portions of the eastern and western United States and in other tectonically active terranes. These processes sculpt landforms, produce sediment, and endanger human life and property. The effect of slope movements is felt not only in the area of the moving hillslopes, but also where the movement encroaches on river channels, changing upstream and downstream conditions.

Slope movements have a great variety of effects on river channels. Most effects are subtle catching the eye only of the geomorphologlst. A few are quite newsworthy, particularly where rivers are dammed, causing flooding in upstream areas, such as at the Thistle (1982, Utah USA), Nakayama (1889, Japan), and Kamenose Landslides (1931-1932, Japan). Of greater hazard, because of the rapid rise of water, is flooding downstream after failure of a landslide dam, as experienced at Nagano City on the Sai River. Japan, in 1847. Less dramatic effects of slope movement include chronic production of sediment and damage to fisn habitat by altering both channel structure and streamside vegetation. High sediment production and channel instability during floods in part reflects sediment delivery by slope movement since the previous major flood.

Little research has been conducted on the effects of slope movement on channels. Systematic observations of channel conditions in the vicinity Of slope movements with known histories are especially rare, because both hillslope and fluvial processes are involved wnich are traditionally studied separately. In this paper we draw on a few field examples from the northwestern United States and Japan to develop a general scheme for classifying and analyzing effects of a range of slope movement types on channels. This analysis has several implications for designing future research and for dealing with hazards from slope movements. In this paper we do not examine several important slope movement-channel interactions, including effects of channel change on slope movement rate and the case where debris slides and avalanches originate in 0-order channels (Tsukamoto et a1.,1982) or hollows (Dietrich and Dunne.1978) and proceed down channels as debris flows.

TYPES OF EFFECTS OF SLOPE MOVEMENT ON RIVERS

The broad range in types of slope movement effects on rivers can be considered in a simple analysis of the balance between rate of material delivery by slope movement and the abilitv of the fluvial system to remove that material. We begin by analyzing these general factors, recognizing that many more specific variables are involved, such as size distribution of the colluvium and location of the failure plane relative to the river channel. Some of the major types of channel responses to colluvium delivery and removal are shown schematically in Figure 1. Rate of delivery of colluvium by slope processes can be expressed as discharge of material per meter of channel length per unit of time. We have estimated these discharges (Figure 1), using typical thicknesses and velocities of slope movements observed in the northwestern United States. The rate of deposit removal depends on both stream power, or competence, and the size distribution of the material.

Slope movements can reduce the width of a valley floor and channel and can raise river-bed elevation. Where the rate of slope movement into a channel is very slow relative to the ability of the river to remove material, there may be little change in channel elevation and

RATE OF COLLUVIUM DELIVERY
BY SLOPE MOVEMENTS
(Discharge per meter of channel length)

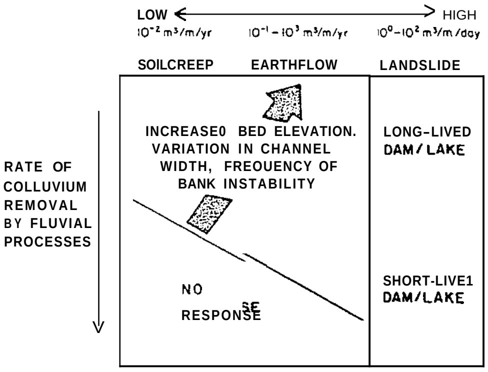


Figure 1. Some responses of channels in relation to rate of colluvium delivery from hillslopes and rate of colluvium removal. Processes and rates of colluvium discharge by slope movements typical in the northwestern United States are shown.

geometry. In this case the slope movement simply produces sediment with minor effect on channel form. Slope movement at a very slow to slow rate (0.06 m/yr to 1.5 m/mo, using the movement scale of Varnes.1978) may exceed the rate of removal by fluvial erosion for periods of up to some years, resulting in temporary channel constriction and an increase in bed elevation. In the extreme case, very rapid movement (greater than 0.3 m/min) of a large volume of material can completely bury a channel and valley floor, forming a dam by locally and abruptly raising the elevation of the valley floor.

EXAMPLES OF EFFECTS OF SLOPE MOVEXENTS ON RIVERS

Extremely Slow Slope Movement

Extremely slow slope movement (less than 0.06 m/yr) occurs predomlnently as soil and rock creep. Typical movement rates neasured in steep lands of the Pacific Northwest of the United States are less than 0.02 m/yr (Swanston,l981). A rate of creep movement of 0.01 m/yr would account for a 1% annual constriction ratio along a 2 n wide channel (1 cm of constriction from eacn side of the channel). Along very small streams this movement could result in localized areas of bank encroachment into the channel. To our knowledge, constriction of mall channels has never been documented, althougn It could be measured with simple techniques, such as repeated surveys of the distance between points monumented on opposite banks of a channel.

Very Slow to Slow Slope Movements

Earthflows (Varnes,1978 , Keefer and Johnson,l984) and other very slow to slow (0.06 m/yr to 1.5 m/mo) slope-movement features encroach on channels on a very significant scale in the Pacific Northwest (Table I). The proportion of channel length with adjacent slope movement features is similar in some geologic terranes of Japan (National Research Center for Disaster Prevention, 1982).

Table 1. Percent of channel length bordered by active and inactive slumps and earthflows on selected rivers in northwestern United States

Location	Slumps/Earthflows Active	Inactive
Redwood Creek California 1/	19	25
Lookout Creek Oregon 2/	a	26
Middle Santiam River, Oregon 3/	25	33

Sources:
1/Nolan et al.,1976. for Highway 299 to Lacks Creek.
2/Swanson and James,l975
3/Hicks,198 2

The range of earthflow-channel relations can be observed in a group of earthflows for which we have some detailed records of movement and channel change. The 17 ha Lookout Creek Earthflow, located in volcaniclastic cocks of the western Cascades, Oregon, flows into Lookout Creek. Channel encroachment is indicated by the 17 m channel 'width outside the earthflow-affected area and a width of only 13 m along the earthflow toe, based on 1977 measurements. Measurement of earthflow movement in 1974-1984 indicates that the toe is advancing into the channel at an average rate of 0.1 m/yr (Swanson et a1,1980), so the channel constriction ratio is about 0.6% of channel width/yr. We assert that sediment production from earthflow toes with such low constriction ratios is episodic. Analysis of landforms and tree ages reveals that major erosion of the Lookout Creek earthflow toe, mainly as streamside debris avalanches, occurred in the winter of 1964-1965 when a maioc flood with approximately 100-year return period took place (Swanson and Swanston. 1977). There have been no additional debris avalanches or significant erosion by other processes at the toe in the 20 years since that flood, but continued eacthflow movement may lead to repeated episodes of debris avalanching with a frequency on the timescale of decades.

However, the valley floor at this site has been constricted by approximately 50 m in the vicinity of the earthflow, based on comparison with upstream and downstream areas (Swanson and Swanston, l977). The longitudinal profile of Lookout Creek through the earthflow constricted section is nearly straight, sugqestinq that the earthflow may have locally raised channel elevation, but not enough to create a convex longitudinal profile (Swanson and Swanston, l977)

The Jude Creek Earthflow also occurs in altered volcaniclastic rocks of the western Cascades, Oregon. The toe of this earthflow has been moving into Jude Creek at approximately 12 m/yr over the 1979-1984 period of observation. At the point of contact with this 30 ha earthflow, Jude Creek is about 6 m wide and drains 800 ha. The rate of channel constriction by this earthflow is therefore about 200%/yr. This relatively high rate of movement into a small channel has had profound effects on the longitudinal profile and channel bank conditions. Numerous small-scale (1 to 100 m^3) debris avalanches and slumps deliver sediment to the channel during each fall-winter period of earthflow movement and high stream flow. The earthflow has raised channel elevation, resulting in deposition of a Low-gradient flat upstream of the earthflow and a very steep stream section over boulders and logs transported into the channel by the earthflow at its Lower end. The result is a stair-step longitudinal profile of the channel through the earthflow-constricted section.

Six earthflows of intermediate constriction ratios have been examined in the Van Duzen River basin, northern California, USA, by Kelsey (1977, 1978). Toes of these earthflows moved at an average rate of 2 to 7 m/yr over the period of 1941 to 1975, based on displacement of distinctive features observed in aerial pnotographs. The channel is 40 to 50 m wide in tne vicinity of tnese earthflows, so they exhibit constriction ratios of about 5 to l0%/yr.. Significant erosion of the toes of tnese earthflows occurred every few years; in some years tnere was no toe erosion (Kelsey, personal communication!. Individual earthflows in this area did not have significant effect on tne longitudinal profile or valley floor width, oecause the river was large enough to remove earthflow colluvium before the riverbed could be raised.

Where Large earthflows have entered fourth-ana fifth-order channels of lower gradient, broaa floodplain and terrace complexes are uevetoped upstream. Examples of such broad valley floors above eartnflows are widespread in the Tertiary volcanic terranes of the western Cascades, Oregon, where eartnflows 50 ha or larger are common. A 100 ha unnamed earthflow dating from more than 6.600 years B.P. (Swanson and James,1975) has constricted Lookout Creek at a point where it.drains 5600 ha. The resulting increase in river bed elevation has contributed to formation of a valley floor up to 300 m wide extenoing upstream for 3.5 km. The valley floor outside tne zone of influence of this earthflow is siynificantly narrower. The development of such features in smaller channels is limited by steep gradients and narrow valleys. Earthflows entering sixth-order and larger channels are Likely to have little effect on long-term riverbed elevation, because these channels have sufficient stream power to remove earthflow deposits (Kelsey, 1977).

We know of no examples in the Pacific Northwest of the United States of failure planes extending beneath a river and emerging at the surface beyond the far river bank. This type of slope movement feature has been observed in Japan, with notable examples including the Kamenose, Shorinzan, and Kujimidai (Komoro City) landslides. In these cases the entire riverbed is lifted up, landslide colluvium may be exposed in the riverbed by scour of alluvium, and a bulge of earth may occur om the far side of the river. The river banks in this situation are likely to be more stable than in the case of earthflow movement directly into a channel that results in channel constriction and widespread bank erosion.

In summary, earthflows can constrict channels where the zone of failure enters a channel. Sediment production from the earthflow toe can be by debris avalanches. Earthflows with a higher ratio of channel constriction (expressed as movement rate of earthflow toe/channel width in units of %/yr) may produce sediment by more frequent, smaller debris avalanches than sites with lower channel constriction ratios. In the latter case, sediment production by debris avalanches from earthflow toes may take place predominently during major floods after some years of slow channel constriction.

However, at the toes of earthflows with high constriction ratios fluvial transport cannot keep pace with delivery of colluvium by the earthflow. particularly where the earthflow transports a significant component of very large boulders (Kelsey,1978). As a result of this local increase in base level, landforms such as broad, low-gradient valley floors can develop upstream of earthflow constrictions and may persist on the time scale of thousands of years in some cases (Swanson and James.1975).

Rapid Slope Movements

Deposits from large, rapid (1.5 m/day to 0.3 m/min) to extremely rapid (.greater than 3 m/s) slope movements can accumulate in a channel and adjacent valley floor, forming a dam. A lake is typically impounded above the dam. Such a dam may eventually fail as a result of piping, collapse by mass movement, or channel incision after overtopping. Many landslide dams do not fail or may be only partially eroded before the channel crossing them is armored by a lag concentration of coarse particles. The associated lakes may persist Eor thousands of years until filled with sediment from upstream areas. The likelihood of failure of landslide dams depends on charactersitics of both the dam and the river. Critical landslide characteristics affecting dam stability include the volume and size distribution of material moved into the channel and the geometry and internal structure of the resulting dam. Channel conditions that determine the ability of fluvial processes to erode the dam include the flow regime, gradient of the channel on the face of the dam, and grain-size distribution of bed materials.

As a first approximation, we hypothesize that, for a particular set of characteristics of dam material, larger dams blocking smaller rivers will have greater probability of forming a lake which persists until filled with sediment, rather than failing catastrophically. On the other hand, short-lived dams (dams which fail) should occur most commonly where smaller landslides enter rivers with larger drainage areas. As a test of this hypothesis, we compare landslide volume with drainage area above the landslide dam (as a surrogate for river discharge), using nine examples from Japan (Figure 2, Table 11). We have limited our analysis to these two variables, because many of the other types of crucial data on river and landslide characteristics are not available from published reports and maps.

significantly influenced by human activities. Excavation of landslide debris at several sites clearly increased the rate of river channel cutting through landslide dams, such as the Wada and Kamenose Landslides. However. field conditions at these sites make it very unlikely that landslide lakes could persist very long under natural conditions. On the Other hand, control structures have been built on the outlets of lakes formed by the Bandai Landslide

Table 11. Examples of landslide-dammed lakes in Japan. Footnotes indicate source of data on landslide volume

Names of Landslide, Lake, River	Landslide Volume (m^3)	Drainage Area Above Dam (km^2)	Lifetime of Lake
Bandai landslide Lake Hibara (1)	1.5×10^9	98	Long-term (1388-present)
Bandai landslide Lake Onogawa (1)	1.5×10^9	40	Long-term (1888-present)
Banaai landslide Lake Akimoto (1)	1.5×10^9	110	Long-term (1838-present)
Yanakubo Pond (2)	0.7×10^6	2.8	Long-term (1847-present)
Kamenose landslide Yamato River (3)	22×10^6	780	Short-term (8 mo)
Kokuzo landslide Sal River (2)	2×10^6	2600	Short-term (19 days)
Hime River Hieaa landslide (4)	3×10^6	460	Short-term (1 mo.)
Wada landslide Niyu River (5)	0.7×10^6	140	Short-term (5 days)
Nakayama landslide Totsu River (6)	3×10^6	510	Short.-te m (1 day)

(1) Ui,1983
(2) N. Oyagi,estimated from map
(3) Ministry of Construction,1980
(4) M. Watanabe,personal communication
(5) Fugita et a1.,1983
(6) Estimated from map in Kagose,1976

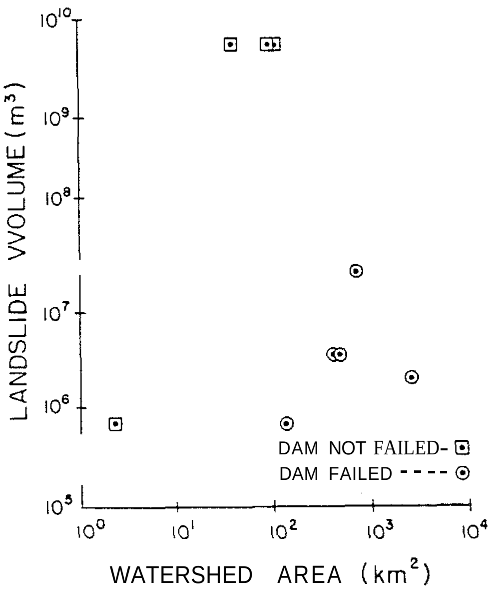
To test the hypothesis, we examine the degree of clustering of Short- and long-lived lakes in a plot of slide volume against dramage area above the dam (Figure 2, Table II). As hypothesized, a significant clustering is observed with a pattern of smaller landslides into larger watersneds being more likely to fail soon after dam formation. Although there are only nine widely scattered points in Figure 2, the pattern suggests that it should be possible to identify combinations of landslide volume and watershed area where catastropnic dam failure would be predicted.

This analysis is limited by the small sample size, tne failure to analyze other variables such as size distribution of landslide debris and river discharge, and the possiblity that the lifetimes of some landslide dams were

which we identify as long-term lakes. Tk is reasonable to assume that these lakes would persist under strictly natural conditions, just as Lake Inawashiro and Spirit Lake (Mount St. Helens, USA) have in identical situations.

An additional complication in analyzing relations among landslide volume, watershed area, and life time of lakes is that, in the case of a very large slope movement over a landscape of low to moderate relief, only a fraction of the deposit may act as a dam. This effect is most pronounced in situations such as at Bandai in 1888 and Mount St. Helens in 1930 where the landslide deposits spread along many kilometers of the valley floor, damming mouths of tributary streams. In these cases a large volume of landslide debris upstream of the tributary may play no role in damming. In Table II and Figure 2, values of total deposit volume are used, which may have the effect of increasing the scatter of points.

Furthermore, analysis of factors such as channel gradient and geologic terrane as it influences size distribution of landslide debris



Fiyure 2. Distribution of long- and short-term landslide-dammed lakes in relation to landslide volume and watershed area.

will be essential to understanding conditions controlling landslide dam failure.

IMPLICATIONS

The geomorpnic observations and hypotheses in this paper have several implications for prediction anti mitigation of sedimentation and nazards to life and property.

The timing and approximate magnitude of sediment production from toes of earthflows can oe predicted from anowledge of channel constriction ratios and time since previous major episode of toe erosion. Sediment production, stream bank instability, and changes in cnannel geometry at earthflow toes may be greatest in a flood after an extended period of years with no large floods.

Bank staoilitation works on the toes of active slope movements, such as earthflows, can eventually decrease overall stability of the site, if channei constriction is occurring at a significant rate. By preventing more frequent, Low-magnitnde erosion of earthflow toes, the potential for erosional events of greater magnitude is ennanced. In this perspective, control measures snould begin on the slope movement itself :atner tnan at the point of fluvial erosion of :ne toe.

The potential for occurrence of floods from release of lake waters as a result of failure of landslide dams can be predicted by zoning watersheds in terms of potentials for producing landslides large enough to form dams at points in the river system where dam failure is likely. Analysis of geology and history of slope movements would yield data on the size distribution of landslides in the past and the probable geographic distribution of large landslides in the future. Analysis of topography would reveal locations in the drainage system where these large landslides are likely to form dams. Analysis of the history of stability of landslide dams in relation to location in the drainage basin (as in Figure 2) or analysis of dam stability using mathematical models such as BREACH (Fread,1984) could then be used to identify sections of the drainage network where landslide dams with significant probability of failure and downstream areas of flooding are likely to occur.

Clearly the analysis of channel response to slope movements is limited by lack of systematic observations of channel changes at sites where slope movement is also monitored. Consequently, we see a need for such observations at sites spanning a broad range of consteiction ratios and landslide volumes in a range of channel sizes. This will require innovative, cooperative efforts by hillslope and fluvial engineers and geomorphologists.

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