M2 Environmental Services
Answers in Geomorphology and Land Use Planning

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Noel Gilbrough
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P.O. Box 3755
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Dear Noel,

Attached is the draft report on geomorphology of the Hazel and Gold Basin landslides. Please review it and call me with any questions and comments. I am also sending a copy to Tracy Drury and to Bob King at HDR Engineering.

Sincerely,

Daniel Miller

Site Descriptions

Hazel

Location

The Hazel landslide is located adjacent to the North Fork of the Stillaguamish River, on the north side, near river mile 20 (T32N, R07E, S1&2).

Materials

The Hazel landslide occurs within a deep deposit of unconsolidated sands and underlying lacustrine (lake-deposited) silts exposed along the margins of Whitman bench, a remnant of the large glacial terrace formed after the last advance of continental ice into Puget Sound. Surface exposures along the landslide indicate a relatively uniform stratigraphy consisting of about 50 meters (160 feet) of well-graded sand overlying an unknown depth of horizontally-bedded silt. Interbeds of finer material and infilled channels indicate that the sand was fluvially deposited. Lack of overlying till suggests the sand is a recessional outwash feature, as mapped by Tabor and others (1988).

Lacking any particle-binding cement, both the sand and underlying silts have limited long-term strength. The sand is unconsolidated and exhibits no cohesion. Measurements of similar deposits elsewhere in Puget Sound (e.g., Palladino and Peck, 1972) indicate some degree of cohesion in undisturbed silts, on the order of tens of kilopascals. This is sufficient for maintenance of relatively steep slopes, but insufficient for creation of relief greater than several tens of feet. Hence, channel incision into these deposits will initiate landsliding. Once disturbed, these materials maintain a residual strength that is considerably lower.

The sands are very permeable, and form an unconfined aquifer overlying the considerably less permeable clay. Groundwater seeps are found at the sand-clay contact. Both the sand and the silt are subject to reductions in strength caused by water saturation of pore spaces. This reduction is proportional to the pore-pressure gradient: increases in watertable elevation can lead to reductions in slope stability. The sand is generally well drained, because of its high permeability, thus rendering slopes underlain by sand to be relatively insensitive to saturation-induced landsliding. However, the underlying silts, because of their very low permeability, cause groundwater to collect in the overlying sand. Thus the sand is in many locations saturated for some depth above the contact. This zone is subject to saturation-induced slumping and seepage induced liquifaction. Where exposed in headscarps and stream banks, the silt deposits appear to be generally saturated. Headscarp silt exposures suggest that slip surfaces on individual blocks within the landslide complex extend below the sand-silt contact. Seasonal movement on many of these blocks indicate that the silts too are subject to saturation induced slumping and respond to variations in load and pore pressure caused by changes in the depth of saturation in overlying sand.

Topography

The Stillaguamish has moved an immense volume of material to create the current valley topography. At Hazel, post-glacial fluvial erosion has cut nearly 200 meters (650 feet) through the valley-filling silts and outwash sands. Incision of the valley initiated large-scale slumping of the deposits along the valley margins. Morphology indicative of large slumps is common along both sides of the river (e.g., mapping by Tabor et al., 1988). Figure 1 shows valley topography in the vicinity of the Hazel landslide, with several large slump blocks along the margin of Whitman bench highlighted. In most cases, these large slumps appear to be currently stable, or at least dormant (Thorsen, 1996). Nevertheless, conditions conducive to mass wasting persist. The headscarps and toe slopes of the ancient slumps themselves commonly present locally steeper slope gradients along which modern landsliding tends to be concentrated. As seen in Figure 1, the Hazel landslide occurs within an old, larger slump.

Landsliding in the glacial deposits also tends to be associated with river erosion into slope toes along meander bends. Thorsen (1996) identified four large, active deep-seated landslides in glacial deposits within the Hazel watershed administrative unit, including the Hazel slide. The common factor in each case was active river erosion of the slope toe.

Since pore pressures influence effective material strength and consequent slope stability, the pattern of groundwater flow presents another major control on landslide activity. Many factors affect the movement of groundwater, including surface topography, the geometry and distribution of subsurface deposits, and the spatial and temporal patterns of recharge by rainfall and snowmelt. Surface exposures of the sand-silt contact along Whitman bench indicate a relatively uniform stratigraphy near the Hazel landslide, so groundwater flow is dictated here in large part by surface topography. Assuming that the sand layer forms an unconfined aquifer, groundwater collecting from recharge over Whitman bench must flow laterally to surface seeps feeding channels along its northeast and southern margins. Active seepage is found year round from banks along Rollings creek to the northeast and into channels and ponds to the west of the Hazel landslide. Two small channels draining the body of the Hazel landslide are also fed year round by seepage from groundwater. These observations suggest a simple aquifer geometry, with groundwater recharge to the Hazel slide coming from a relatively limited zone extending onto the Whitman bench to the northwest and into the Headache Creek basin to the north (Benda et al., 1988; Miller and Sias, 1997).

History

The Hazel landslide has been active for over half a century. Thorsen (1996) noted a tight river bend impinging on the north bank with active landsliding visible in 1937 aerial photographs. The next 60 years involves two periods of relatively low landslide activity, and two periods of relatively high activity, the last of which extends to this day.

Benda et al. (1988) estimated a landslide size of approximately 10 acres from 1942 photographs. Tracings made by Curtis (1988) show relatively little activity at that time. The landslide appears relatively quiescent in 1947 photographs (Figure 2). In 1951, however, a mud flow from one of the channels draining the landslide briefly blocked the river. In 1952, Shannon and Associates reported activity extending up to 1000 feet from the river's edge, with headscarps 70 feet high and downslope motion of large, intact blocks. They also reported persistent turbidity from channels draining the landslide. The photographs shown in Figure 2 indicate that landslide activity persisted over the next decade, with the river at the toe throughout this period. A berm of alluvium and logs was constructed in the summer of 1960 in an attempt to protect the toe of the landslide. It was eroded away the first winter (Webb and Rees, 1961). A rock revetment was constructed along the eastern portion of the slide toe in 1962, which persists to this day.

Despite protection of the toe by the revetment, a large event occurred in January 1967, which dramatically altered the geometry of the landslide and the river channel (Figure 3). Downslope translation of a large block and accompanying mud flows pushed the river channel about 700 hundred feet to the south. Debris from this event then insulated the landslide toe from further bank erosion. Over the next 20 years, gradual revegetation of the landslide scar marks a period of relative quiescence. Several persistent scarps, however, indicate continued adjustments within the body of the landslide.

Throughout this time, the river was gradually eroding through debris at the toe, moving northward at a rate of 30 to 50 feet per year (Wolff, 1988). By 1988 it had reached the toe, and an event in November of that year again forced the river south. Debris volume was considerably less than that involved in the 1967 event, and the river soon regained its former channel. The 1991 and 1995 photographs show scarps and mudflows along the western portion of the landslide. These are eating into a relatively intact block that was last displaced during the 1967 event. Observations from across the river in 1999 indicate little additional headward expansion within this zone since the summer of 1997. The 1962 revetment has been re-exposed, at least since 1996, along the eastern, so far inactive portion of the slide.

Geomorphology

Mass Wasting

Landsliding at Hazel has involved four mechanisms of mass-movement:

1) <u>Translational motion of large blocks</u>. Such movement can be sudden, fast, and over a large scale. The 1967 event provides a spectacular example. A large block, carrying a section of forest with trees standing, slid from near the headscarp down to the river, traversing a slope distance of nearly 800 feet. Slow, incremental movement of large blocks also occurs, manifest by the persistence of headscarps, development and enlargement of tension cracks, and gradual displacement of trees.
   Controlling Factors: Although river erosion has created topography conducive to this type of activity, motion of the blocks themselves seems to be triggered primarily by variations in pore pressure and the flux of ground and surface water from the landslide itself (Miller and Sias, 1997). For example, the well-traveled 1967 block was initially about 400 feet from the channel: the river had no access to the block directly. Likewise, seasonal activity on upslope blocks points to movement in response to winter increases in groundwater flux.

2) <u>Slumping along steep, stream-cut banks</u>. Slumping has occurred at a fairly large scale, involving thousands of cubic meters of material, along river-cut banks of the Stillaguamish. It occurs at a smaller scale, involving tens to hundreds of cubic meters, along channels draining the body of the landslide. Catastrophic movement of such slumps has resulted in disintegration of the blocks and development of debris flows. In other cases, movement is incremental. The block shifts to a stable position without disintegrating.
   Controlling Factors: These events seem to be triggered directly by stream erosion at slope toes.

3) <u>Headward-migrating slumping</u> along small channels draining the body of the landslide. The enlargement of the active area seen between the 1991 and 1995 photographs involves this type of mechanism.
   Controlling Factors: Two factors drive this type of activity. Seepage, typically emanating near the silt contact, erodes material from slope face, causing steepening and undercutting of the slope and slumping of overlying material. High rates of surface discharge can also rapidly incise the small channels, causing rapid slumping of bank material. Both mechanisms result in upslope migration of a knickpoint. Failed material typically disintegrates and is flushed out through the channel.

4) <u>Earthflow (mudflow) movement</u>, both into the Stillaguamish directly and into channels draining the body of the landslide. Such flows are typically created by disintegration of slump

blocks. Catastrophic slumping has initiated rapidly moving debris flows. Slow or incremental slumps have evolved into persistent flows, that provide a chronic source of fine sediment both to the Stillaguamish directly and to channels draining into the Stillaguamish.
Controlling Factors: Rates of earthflow movement are probably controlled primarily by water content and variations in pore pressure.

Two primary factors act to drive motion of the landslide: 1) changes in geometry caused by fluvial erosion, both along the banks of the Stillaguamish, and in channels draining the body of the landslide, and 2) variations in pore pressures and water flux (both groundwater and surface water) to the landslide. We might therefore anticipate that activity on the landslide will correlate with seasonal precipitation volumes and variations in river discharge.

Figure 4 shows summary climate data from weather stations at Arlington and Darrington. Hazel falls about midway between these two stations. The upper two graphs show cumulative annual precipitation depths, integrated over the water year (October 1 of the previous year through September 30). Total precipitation is greater at Darrington, but year to year variability tends to correlate between the two sites, i.e., wet years at Arlington are also wet years at Darrington. Arrows indicate the timing of the three major, river-blocking events on the Hazel slide, those in December 1951 (water year 1952), January 1967, and November 1988 (water year 1989). Numbers above the arrows indicate the percentage of the years of record having cumulative precipitation greater than that year. These graphs show no clear connection between cumulative annual precipitation and the occurrence of major events.

We may also examine multiyear patterns. The decade prior to 1951 was a period of relatively low landslide activity. This is also a period with an extended run of below average rainfall. Besides that, however, no clear relationships are discerned from these graphs. The 1967 event follows a pair of below average years; the 1988 event follows a string of three below average years.

It is also possible that individual storm events are triggering landslide activity. The bottom graph shows the peak annual 5-day accumulated rainfall depth at the Darrington station. Although major events are associated with the largest storms of the year, they are not occurring in the years with the largest storms. These observations reveal no clear connection between temporal patterns of precipitation and landslide activity.

We can also look for relationships between discharge in the Stillaguamish and activity on the landslide. Figure 5 shows both cumulative and peak annual discharge at the gage on the North Fork near Arlington. This is the only gage in the basin with a long, continuous record. I plot cumulative discharge as a measure of total transport potential. Since sediments composing the landslide are predominately fine grained, even relatively small flows may move material delivered to the channel. Peak annual discharge provides a relative measure of the erosive potential of individual flood events. Again, although the largest landslide events tend to occur during high water, coincident with large storms of course, there is no clear relationship between the largest events and initiation of landslide activity.

The mixture of processes and controlling factors involved in mass wasting at Hazel make it impossible to identify single triggering events responsible for initiation of landslide movement. Activity on the landslide involves a series of interacting processes, with one set of events setting the stage for subsequent series of events. Over the period of record, the largest factor appearing to limit activity on the landslide is insulation of the toe from erosion by the Stillaguamish river. Debris from the 1967 event displaced the river southward. It took 20 years for the channel to regain its former position at the toe of the landslide. Over that time, the landslide revegetated with no signs of major movement. Activity within the landslide reinitiated only after the river again impinged on the toe. Certain types of activity on the landslide do respond to seasonal fluctuations in groundwater flux to the slide, but long-term maintenance of conditions conducive to such activity appear to require continual removal of failed material from the landslide toe.

Delivery

Delivery of sediment to the Stillaguamish occurs in three ways:

1) Direct deposition of mass wasting debris into the river.
2) Fluvial erosion of previously deposited landslide debris and of banks at the toe of the landslide.
3) Fluvial transport of sediment by small channels draining the body of the landslide.

The first and second probably account for the largest volumetric flux of fine sediment into the river; the second and third account for chronic and year-round flux of fine sediment to the river.

Main-Stem Response

The most direct effect on the mainstem has been blocking and diversion of the channel by landslide debris. The January 1967 event displaced the channel about 800 feet southward and caused avulsion of a side channel (see the photographs in Figures 2 and 3). Later photographs show no obvious up- or downstream consequences of this diversion. The greatest subsequent change is gradually northward migration of the channel through the deposited debris, back towards its former channel, and abandonment of the side channel after several years.

It is not that reaches up and downstream of the landslide are unchanged over time. In fact, the photographs show gradual changes in active channel width, bar size, and channel migration. These changes do not appear to correlate, however, with diversion of the channel by the landslide, or with rates of landslide activity.

The lack of channel plan-form response to the landslide may be due in part to the nature of the landslide debris. Sediment shed from the landslide is predominately fine grained, and incorporated into the wash and suspended load of the river. The channel here is predominately coarse-grained alluvium, transported as bedload from upstream. Channel change is caused by erosion or deposition of these coarser, bed-forming sediments. Thus introduction of bedload from Rollins creek, for example, may have had a larger effect on bar size and active channel width than introduction of fine sediment from the landslide.

Future Activity

The descriptions above lead to certain inferences about the future of the Hazel landslide. There is a strong tendency for the Stillaguamish to maintain its position at the base of the landslide. Despite several minor blockages (1951, 1988, 1996) and one major diversion (1967), the channel has persistently moved back up to the landslide toe. The river thus efficiently removes all material deposited at the base of the landslide and erodes the landslide toe. This activity maintains steep, unstable scarps at the base of the landslide, which promotes persistent slumping into the river.

The gradual changes in landslide geometry resulting from failures at the toe can reduce stability of the entire slide mass (Miller and Sias, 1997), thereby increasing rates of activity upslope and increasing the potential for a large catastrophic failure. The potential for headward migration into the Headache creek basin has been suggested (Shannon and Associates, 1952, Benda et al., 1988). Capture of Headache creek could dramatically increase the flux of water to the landslide. Headscarp advance towards Headache Creek has been negligible since 1952. The 1967 event apparently created a relatively stable geometry over the central body of the landslide. However, analysis by Miller and Sias (1997) indicate a small potential reduction in stability over an area extending into the Headache Creek drainage in response both to erosion of the landslide toe and to incision of channels draining the landslide. Continuing alteration of the toe geometry by the river, and continuing changes to channel geometry over the body of the landslide, may result in changes in landslide behavior and renewed initiation of progressive, headward slumping to the north.

The potential for large-scale destabilization also exists over the western portion of the landslide. Current activity is concentrated along the western portion of the toe with headscarp advance into a large, old, downdropped slump block. This block sits at the base of Whitman bench, buttressing a slope with over 200 feet of additional relief to the top. Gradual reduction of this block by mass wasting reduces loading of the toe of this slope, with the potential for triggering renewed slumping of Whitman bench.

The statements above are not predictions, but inferences based on observations described above and on analyses described in previous reports (Miller and Sias, 1997). I currently have no basis for estimating the probable rate or timing of future landslide activity. The primary conclusion to be drawn is that mass wasting activity will persist for as long as the river remains at the toe of the landslide.

Gold Basin

Location

The Gold Basin landslide is located adjacent to the Sorth Fork of the Stillaguamish River, on the north side, between river miles 32 and 33 (T30N, R08E, S13,14&24). It is across the river from the U.S. Forest Service Gold Basin campground.

Materials

Like Hazel, the Gold Basin landslide also involves glacial sediments. However, its mode of operation is quite different. The Gold Basin landslide involves a complex of three separate, headward expanding lobes. Differences in behavior between Hazel and Gold Basin result, in part, because of differences in the materials involved.

The glacial deposits at Gold Basin are also comprised primarily of sand and silt. However, their spatial organization is different. The stratigraphy of the in situ deposits at Hazel is quite uniform, compared to those at Gold Basin. The Gold Basin deposits are more heterogeneous, both vertically and laterally (see stratigraphic profiles in Benda and Collins, 1992). Thick sequences of sand are exposed within scarps lining the landslide and adjacent to the Stillaguamish, but with interbeds of both coarser and finer material. Laterally, sand deposits grade abruptly into thick sequences of horizontally bedded silt. Drop stones, some quite large, are abundant within these silts. These attributes are characteristic of ice contact deposits (Booth, 1989).

The spatial heterogeneity in the juxtaposition of permeable (sand) and impermeable (silt) deposits results in complex and unpredictable patterns of groundwater flow. Seeepage from the banks indicates many zones of perched groundwater. In places, in-situ silt deposits extend to the ground surface, resulting in local areas with low infiltration capacity and the potential for generating overland flow.

Topography

Regional topography is shown by the shaded relief image in Figure 6. The three lobes of the Gold Basin landslide lie within a larger, bowl-shaped basin draining to the Stillaguamish. Total relief varies from about 160 m (520 feet) along the upstream-most lobe (number 1 in Figure 6), to 220 m (720 feet) along the downstream-most lobe (number 1 in Figure 6).

Topography leading to the Gold Basin landslides is convergent over several spatial scales. Such convergence can cause topographic concentration of groundwater flow. At the largest scale, the bowl-shaped depression directs water flow toward the Stillaguamish valley. Each landslide lope lies within smaller-scale basins overprinted on the major slope leading from the ridge crest to the river. Topographic convergence defining these basins tends to focus water flow toward the sub-basin axis. Note the presence of similar sub-basins adjacent to the landslides: a smaller one in the downstream direction and one of equivalent size upstream. In contrast to the Whitman bench upslope of the Hazel landslide, upslope areas of groundwater recharge here are relatively steep. Groundwater response to precipitation is probably quicker here, with greater sensitivity to individual storm events.

History

Landslide activity has varied over time both within individual lobes and between lobes, as shown for the period of available aerial photography in Figure 7. Additionally, Benda and Collins (1992) show tracings of activity extending back to 1942. The active area in 1942 involved only about one quarter of that in 1964, and was confined to the lower portions of the two upstream lobes. By 1964, the upstream lobes have extended substantially headward and have attained a size that increases little in subsequent years. A substantial fan has formed at the mouth of the upstream-most lobe. Also seen in the 1964 photographs is a zone of active mass wasting just downstream of the middle lobe, between it and the downstream lobe. Although the headscarps surrounding this area persist, there is no further headward expansion.

The period from 1964 to 1972 involved slight headward growth to the east of the upstream lobe (#3 in Figure 6), and revegetation of the fans at the base of the lobes. By 1976, the downstream lobe has become active, the middle lobe has grown slightly to the north, and growth of the upstream lobe has ceased. Note also the landslide scar upslope of the downstream lobe. The downstream lobe continues to expand through 1983, with only minor changes to the upstream lobes. Activity through the early 1990s is relatively minor, with slight headward growth of the middle lobe.

The late 90s have seen substantial reactivation of landslide activity within the downstream two lobes. Headward expansion has occurred both to the north and west on the middle lobe, and toward the northeast on the downstream lobe. Channel-lining vegetation through the middle lobe has been swept away by debris flows traversing the channel. A suite of new debris-flow terraces provides evidence of multiple events spanning the last two or three years, with substantial fluvial erosion of the deposits between mass-wasting events. A large fan has now coalesced from the mouths of the two downstream lobes. This fan has completely filled the channel north of the vegetated island.

Geomorphology

Mass Wasting

Two primary modes of mass wasting are evident at Gold Basin:

1) Headward-migrating slumps. These slumps tend to occur either where seepage along silt contacts is eroding the slope face, or where surface runoff has eroded a gully or channel. Gully erosion through the glacial deposits can be spectacular. Rapid incision has cut slot-like gullies that extend tens of feet vertically through the deposits. Vertical faces are short lived in this material, and further erosion will trigger steep, translational slumps. The material disintegrates as it fails and provides little buttressing support at the base of the slope. The geometry of the exposed scarp is also unstable, thus promoting landslide expansion via headward marching of progressive slump blocks.
   Controlling Factors: Water plays the major role in this style of mass wasting, through seepage erosion, gully erosion, and transport of failed debris from the base of slopes. Thus increases in water flux can result in increased rates of landslide activity. However, since landsliding itself can destabilize upslope areas, mass-wasting activity, once initiated, may persist even without increased water yield.
2) Evolution of landslide debris into debris flows. Silt-rich deposits can evolve into debris flows during landsliding. Debris flows can transport large volumes of material onto and down the sub-basin floor, with direct deposition into the Stillaguamish.
   Controlling Factors: Generation of debris flows on the Gold Basin landslide appears to require silt-rich material. Debris deposits of sand are abundant along the sub-basin valley floors through those areas eroded into clean sand. These deposits tend not to run out into debris flows. Clean sands may be too well drained for creation of a liquid slurry.

Unlike Hazel, these landslides do not appear to be sensitive to bank erosion along the mainstem. The slopes are probably sensitive to conditions at the toe, but in this case, the slopes drain to channels tributary to the Stillaguamish. Fine-grained landslide debris buries the base of the toes: channel erosion appears to play a minor, if any, role in stability of these slopes.

Arguments presented above suggest that the Gold Basin landslides may be responsive to precipitation events. The stratigraphy and topography (i.e., impermeable silts at shallow depths and steep slopes) are conducive to rapid groundwater fluxes. A comparison of the precipitation record shown in Figure 4 supports this view. The smallest extent of recorded landslide activity was in 1942, following a string of two below average years for total precipitation and a string of three years with relatively small storm events. The greatest volumetric flux from the landslide occurred over the period from 1942 to 1964, with major expansion of the two upstream lobes. This same period saw an increase in the frequency of larger storms and of very wet years, starting in 1950. Growth of the downstream lobe between 1972 and 1976 occurred during a period with two of the wettest years of record and with two large storm events, 1974 and 1976. Reactivation of the downstream lobes since 1995 corresponds to a period having very large storms every year.

Delivery

Delivery of sediment to the Stillaguamish occurs though three mechanisms:

1) Direct deposition by debris flow. Debris flows deposit both over the valley floors of the sub-basins and onto fans at the mouths of the sub-basins. The incidence of debris flows has probably increased in the last several years, as the downstream lobes have advanced into a very thick deposit of debris-flow-generating silt.
2) Fluvial erosion of landslide debris by the channels draining each sub-basin. The flux of fluvially transported sediment out of each lobe probably tracks landslide activity. Deposition of debris over the valley floor provides a fresh source of sediment to these channels. Terraces formed by fluvial erosion of debris flow deposits indicate that these channels can move substantial volumes of material. As they gradually flush material from their beds, a lag of

coarser material develops which acts to armor the bed and reduce further bed erosion. The glacial deposits here contain a small amount of cobble and boulder-sized clasts. These have effectively armored much of the bed of the upstream lobe. Similar armoring will occur as activity wanes on the downstream lobes. Woody debris within the channels also acts to inhibit downstream flux of sediment. Where channel-spanning logs are found, the channel exhibits a stepped profile, with stored sediment upstream and a vertical drop over the wood. The source for this wood are the trees captured by headward expansion of the landslides into standing forests. Much of the wood has been carried downstream in debris flows.

3) Erosion of fans formed at the mouths of the lobes by the Stillaguamish river.

Mainstem Response

Development of the vegetated island followed the large pre-1964 flux of sediment from the upstream two lobes. Channel plan form prior to that time is unknown without earlier photographs. Since 1964, the island has been a stable feature with the extent of vegetation increasing over time. Through 1998, channel plan form remained essentially unchanged. Accumulation of a large log jam at the upstream end of the island in the early 1990s caused flow diversion into the south bank, prompting erosion-control measures by the Forest Service to protect the Gold Basin campground.

An increase in the rate of flux from the downstream two lobes in the last several years has caused formation of a large fan, which has filled the channel north of the island and currently forces all flow through the channel south of the island adjacent to the campground.

Future Activity

Past activity is characterized by rapid headward expansion and accompanying large volumetric flux of landslide debris to the Stillaguamish over a period lasting about a decade. This is followed by a period of revegetation over the landslide scar, with minor but persistent headscarp activity. The three lobes do not respond synchronously, but appear to act independently. Landslide activity may be initiated during years with above average precipitation, and may occur in response to large rainstorms. Although the currently available information is insufficient to resolve this issue, it may be that, once initiated, mass wasting activity within an individual lobe varies in response to multiyear patterns of precipitation. The decadal response period may correlate with decadal variations in precipitation.

The last several years have seen renewed activity on the downstream two lobes. These have both now accessed a thick silt deposit, which promotes initiation of debris flows that run out to the mouth at the Stillaguamish. If history is any guide, this activity will persist for several more years, and then gradually decline. Delivery of material to the Stillaguamish will be rapid. Much of the previously available storage volume in the channel and over the fan is currently filled. The channels and fan at the mouth now provide an efficient transport corridor for debris flows directly to the Stillaguamish. Until the flux of mass wasting debris declines, the channels crossing the fan will likely change location each winter, as they incise through new debris flow deposits.

Over the long term, mass wasting flux from the Gold Basin slide will decline as the glacial deposits are depleted.

Strategies for Reducing Sediment Flux to the Stillaguamish

There are two potential means for reducing sediment flux: 1) Slope stabilization to reduce rates of mass wasting, and 2) Stabilization of channels to reduce delivery of mass wasting debris to the mainstem. I'll discuss the each site in turn.

Hazel

Slope Stabilization.

Two means exist for stabilizing the slope: 1) Protection of the toe to prevent erosion into the landslide and to prevent erosion of landslide debris deposited at the base. 2) Reduction of water flux to and over the body of the landslide.

Toe Protection

<u>Revetment</u>*:* The eastern portion of the landslide has been relatively stable since construction of the rock revetment along the toe in 1962. This may be a coincidence, but it does suggest that protection of the toe from further erosion can prevent further destabilization.

Potential Problems: Erosion behind the revetment was occurring along its western end in 1996, indicating the need for monitoring and maintenance of such structures, and the potential that they will fail in long-term protection of the toe. The revetment provides no protection of debris that does fail and does not prevent delivery of failed material to the channel. It was completely overrun during the 1967 event. A revetment also does nothing to stabilize an already destabilized slope, as demonstrated by the failure in 1967.

<u>Channel Diversion</u>*:* The period of relative quiescence for two decades following the 1967 event, during which time the toe was insulated from the river by landslide debris, illustrates the potential effect of toe protection by river diversion. As shown in Figure 4, this period included some of the wettest years on record at the Darrington station (i.e., water years 1971, 1972, 1974, and 1976) and a large storm in water year 1974. Minor activity persisted on the landslide, but there were no long-runout failures and no direct delivery of mass wasting debris to the Stillaguamish.

Potential Problems: There are three primary issues of concern:

1) Initiation of up- and/or downstream channel and bank scour associated with changes in flow pattern caused by the diversion. Historically, the large natural diversion in 1967 did not initiate large-scale channel adjustments up or downstream. Nevertheless, this issue deserves further scrutiny. In particular, downstream areas of potential bank erosion need to be identified. There are structures adjacent to the river just downstream of the landslide. Furthermore, erosion of the slope toe beyond the west end of the landslide could initiate slumping at that location. Any efforts at channel diversion should include protection of the bank along the outside edge of the bend in the channel to the south, downstream of the landslide.

2) Overrunning of the diversion structure by landslide debris. The 1967 event demonstrates the potential for dramatic runout. The change in geometry caused by that event probably resulted in a more stable slope configuration over the central portion of the landslide, but continued activity since that time indicates continued instability. Does the potential for a large, 1967-style, event persist? The methods described in Miller and Sias (1997) can provide an estimate of the volume available for mobilization in a large, catastrophic event.

   Using topography based on 1978 aerial photography, the stability model of Miller (1995) was used to estimate depth to the slip surface over all areas of the landslide at various levels of estimated stability. This model assumes plane strain, limit equilibrium, and a slip surface that may vary from circular to linear. Estimation of material properties and pore pressures is described in Miller and Sias (1997). At any location, multiple slip surfaces can be defined, each with its own balance of forces and a specific estimated factor of safety. In previous analyses, we looked for the least-stable surface. With this analysis, I look at all surfaces having a specific factor of safety value. Even if not the least-stable geometry, any geometry that is unstable is subject to failure. The goal here is to find the largest volume subject to landsliding.

Factor-of-safety estimates obtained with this model provide only relative measures of absolute stability. I don't know precisely what value indicates an unstable slope. So, I've done the analysis for several values, and compare results. Figure 8 shows estimated depths to the slip surface for a factor of safety of 1.0. This produces estimated slip surface geometries consistent with the style and size of slumping occurring on the western, active portion of the landslide since 1990.

That analysis provides an estimate of the volume that could be mobilized in a large, catastrophic slump. Following the transect A-A' shown in Figure 8, the total volume mobilized is 5425 m^3 per meter width. To estimate the runout length, I assume the debris will deposit in a wedge with a 15% surface slope. This geometry is consistent with laterally unconfined landslide and debris flow deposits measured both here and at Gold Basin. The estimated total runout length is 270 m (880 ft). This is very similar to the runout length of debris in the 1967 event.

The analysis described above does not account for progressive failure that may occur as landsliding alters slope geometry. To estimate volumes that could be mobilized by further destabilization, Figure 9 shows estimated slip-surface depth for a factor of safety of 1.3. These results are largely speculative, but they do illustrate the potential consequences of destabilization of slopes leading to Whitman bench, and of slopes to the west of the landslide. Estimated volumes increase by an order of magnitude.

3) Failure of the diversion, and avulsion or migration of the channel to its former position. I have no basis for evaluating the potential for such failure. If it were to occur, the situation would simply revert back to what it is now: the river at the base of the landslide.

Dewatering the Landslide:

There are three options for reducing water flux.

1) Surface water diversions. All surface drainage over the landslide is fed by groundwater seepage from within the landslide, so there is no option for reducing surface flow to the landslide. However, channels draining the landslide could be lined or otherwise stabilized to prevent erosion of their beds and banks. There may also be potential to promote more rapid drainage of wetland areas associated with groundwater seeps.
   Potential Problems: Channels draining the landslide are flowing over landslide debris. In some cases they are riding on top of existing slump blocks. Structures such as drainage ditches are subject to disruption by movements and small adjustments that occur annually over the body of the landslide. In addition, mass wasting into the channels may inundate any structures. The potential for failure of drainage structures is high.

2) Drainage or pumping of subsurface flow to intercept groundwater flow to the landslide or to actually drain subsurface water from the landslide. Such methods are used extensively elsewhere for protection of roads and structures. Use of dewatering methods here would require additional study to determine a feasible and optimal design.
   Potential Problems: Analysis done by Miller and Sias (1997) indicates that groundwater fluctuations play a role in landslide activity. However, that analysis also indicated that the influence of groundwater was considerably less than the influence of river erosion at the toe.

Efforts here, without protection of the toe, would probably be ineffective at stabilization, but may serve to reduce or delay flux of material from the landslide.

Reductions in recharge to groundwater. Recharge varies in response to variations in precipitation and to variations in the amount of water intercepted and evaporated or transpired by vegetation. Miller and Sias (1997) used numerical models to estimate the effects of timber harvest in the groundwater recharge area on landslide behavior. Estimated effects are small, but integrated over time, may substantially influence total volumetric flux of fine sediment from the landslide. The options for reducing recharge, however, are limited. The primary goal may be to prevent increases in recharge (DNR, 1996).
Potential Problems: Again, water plays a secondary role in maintaining landslide activity. Without efforts to prevent erosion of the toe, reductions in groundwater flux will not result in stabilization of the landslide.

Reducing Delivery:

The only option for reducing delivery of mass wasting debris to the Stillaguamish is diversion of the channel. As long as the channel is against the toe of the landslide, there is no area for storage of landslide debris. Likewise, the area available for storage of debris is a direct function of the distance between the channel and the landslide. The larger this distance, the greater are opportunities for storage of sediment and the greater are the likely reductions in flux of sediment to the Stillaguamish. Note, however, that any area provides only a finite storage volume. This volume may be roughly estimated in the same way I estimated runout: assume accumulation of a fan with surface slope of about 15%. Once filled, sediment flux to the river would, on average, equal that off the landslide.

Diversion of the river from the toe should, however, serve also to reduce sediment flux from the landslide. As debris piles up at the toe, the landslide will achieve a more stable configuration, and the rate of mass wasting will decrease over time. One design goal may be to estimate the total volume of sediment available for fan construction, and place the diversion sufficiently far away to accommodate the fan. Estimates of runout made above suggest a fan length of about 900 feet. However, stabilization of the landslide itself, with consequent reductions in mass-wasting sediment flux, should be achieved, even if the debris fan reaches the diversion structure.

Channels draining the landslide will continue to drain to the Stillaguamish following diversion of the mainstem. Because of the fine grain size of sediment shed from the landslide, these channels will continue to carry silt as wash load and deliver turbid water to the river. Reduction of turbidity requires reductions in the introduction of silt to these channels. Increased landslide stability will reduce mass wasting delivery of sediment to these channels, but the unstable nature of the glacial sediments ensures some level of continuing mass wasting. The channels may also incorporate fine wash load through bank erosion and scour of their beds. Hence, increases in channel stability may reduce turbidity, particularly during summer low flow. Channel stability may be increased by armoring of the bed with coarse clasts, vegetation of the banks, and presence of in-channel woody debris.

Recommended Option

Diversion of the mainstem will act both to stabilize the landslide (by protecting the toe) and add storage area for sediment shed from the landslide, which will reduce delivery of sediment to the river. The simple analysis presented above suggests that the diversion should be located to direct the channel course at least 900 feet, at its farthest extent, from the current base of the landslide to accommodate runout of landslide debris.

Further Study

The options described above highlight two issues that require elaboration.

1) Potential upstream and downstream impacts of flow diversion. Diversion of the mainstem will alter flow patterns, with potential consequences for patterns of flood inundation, bank erosion, bed scour, and bar deposition both up and downstream. The diversion accomplished by the landslide in 1967 produced no large-scale channel effects up or downstream that were visible in aerial photographs over subsequent years. Nevertheless, the potential effects of any proposed diversion should be considered. At the least, areas of potential bank erosion downstream should be identified.

2) Total available volume of landslide debris. The analysis presented above is based on outdated topographic data and unverified estimates of watertable geometry. Better volume estimates could be made using up-to-date surveys and with piezometer installations to measure and monitor groundwater fluctuations. Such a study could in fact have applications much broader than to Hazel itself. The role of groundwater on slope stability is still an issue of debate. In particular, the effects of vegetation on rates of recharge are poorly quantified. Data collection and monitoring of the Hazel landslide could serve both to improve estimates of future landslide volume and to improve estimates of slope stability and the effects of landuse region wide.

Gold Basin

Slope Stabilization

There are no obvious means to stabilize the Gold Basin landslide. Toe protection would probably have little, if any, effect on landslide rates. The primary process driving mass wasting appears to be water flux to the landslide complexes. This water is derived from a source area extending at least to the drainage divide. Lack of access and heterogeneous stratigraphy make subsurface dewatering a complex and potentially fruitless exercise. Reductions to recharge through evapotranspiration are probably already maximized, since a mature forest occupies the uplands. The main option is simply to maintain this forest. However, the effects of forest removal and consequent changes in hydrology on the landslide are unknown. Benda and Collins (1992) note expansion of the upstream lobe to the east following clear-cut harvesting of timber upslope of that area. However, the middle lobe also expanded substantially during that time, with no changes in upslope vegetation.

Reducing Delivery

The primary means for reducing sediment flux from the Gold Basin landslides is to reduce delivery of sediment to the main stem. If mass wasting activity continues, as it is expected to, reductions in delivery entail increases in long-term storage along the sub-basin valley floors and in fans at their mouths. I'll discuss each in turn.

Fan Storage

The photographic record shows debris fans at the mouth of each lobe. The volume of these fans has expanded over time in response to inputs from the landslide, and contracted over time in response to erosion by the streams that flow over them and erosion by the Stillaguamish itself. In general, fan volume is low because the high sediment transport capacity of the mainstem keeps pace with delivery from upslope. Over the past two years, however, debris flow transport has caused sediment flux from the landslide to greatly outpace sediment removal by the river. Fans from the downstream two lobes have filled the north branch of the mainstem channel. Since the river is prevented from further displacement to the south by the Forest Service campground, the available fan-storage volume is essentially filled. There are no obvious means of increasing this volume. Hence, sediment from upslope will be carried over the fan surface to the Stillaguamish. The fan may act as a transient storage reservoir, thus modulating sediment output during low-flow years. Ultimately, however, what goes in will come out, so that erosion of the fan will likely increase sediment fluxes to the mainstem in other years.

One option now is to prevent erosion of the current fan. This may have no effect on future, long-term rates of sediment influx from the landslide, but it will prevent downstream transport of the substantial volume (on the order of 120,000 m^3) of sediment currently stored in the fan. Maintenance of the fan would require that the toe be protected from bank erosion by the Stillaguamish, and that the body be protected from erosion by the channels that traverse it.

<u>Protection of the Fan Toe</u>: Engineered structures could be placed to protect the base of the fan from erosion by the Stillaguamish.

Potential Problems: Since the stream draining the middle landslide lobe joins the mainstem here, any structure would also be subject to erosion from behind by this tributary channel. Likewise, the structure would be subject to inundation by debris-flows exiting the middle lobe.

The long-term removal of the water-conveyance capacity provided by the former northern branch of the channel may also pose certain consequences. Stage relationships for the south branch will be altered, with potentially higher flood elevations and/or increased water velocities. This may increase the potential for flooding of the Gold Basin campground. Flooding did occur in the winter of 1998, at which time the north branch may have already been blocked. Alluvial cover in the channel here is very thin, so channel adjustments will probably occur primarily through bank erosion. The log jam that formed at the upstream tip of the island in the early 1990s provides some evidence of the potential effects of changes in water flow paths. Diversion of flow by the jam caused erosion of banks into the campground.

Also likely is the potential for erosion of the north bank into the island. Erosion of the island could increase conveyance capacity of the channel, but migration of the channel through the former island and into the fan would negate the benefits originally sought with upstream bank protection. Vegetation currently on the island may act to minimize channel migration.

Stabilization of Tributary Channels

Streams draining the sub-basins can erode landslide debris from the valley floors and can erode material from the fans. Stabilization of these channels occurs through three processes:

1) Armoring of the bed by coarse clasts. This process occurs naturally as fine sediment is flushed from the bed, leaving a lag of coarser material.
2) Storage of sediment trapped by large woody debris. Woody debris serves two purposes here: it increases hydraulic roughness, thus reducing the sediment transport capacity of the channel, and it acts as small dams that store sediment in upstream wedges. Both the upstream and downstream-most tributary streams currently have woody debris in the channels that is actively storing sediment. The middle tributary is still overwhelmed by debris flow deposition, and wood currently has little effect on the channel. The primary source for wood to these channels are trees from the upland forest carried to the valley floor by landsliding. The deciduous cover in the upstream lobe is now reaching 30 years in age, and is also providing some wood recruitment from riparian areas.

3) Growth of bank-side vegetation. This process offers the primary option for remediation, for both the short and long term. Efforts can be made to encourage rapid revegetation of riparian areas. Additionally, tree plantings could be used to ensure a source of riparian wood recruitment for future years.

Recommended Option

Protection of the fan toe and revegetation of the fan surface and riparian areas along channels draining the landslide lobes. These projects will not reduce future sediment flux from the landslide to the river, but will preclude, or at least delay, introduction of the fan-stored sediment to the river.

Further Study

Permanent closure of the north branch of the channel may have consequences for flooding and erosion on the south side of the river. Rough estimates of flooding potential might be made using a stage rating curve based on a channel survey. In any case, flow depths and channel behavior can be currently monitored, while the north channel branch is blocked.

References

Benda, L., and B. Collins, 1992, *Slope Stability Investigation of the Crown Pacific Property in the South Fork of the Stillaguamish River Basin*, unpublished report, 37 p.

Benda, L., G.W. Thorsen, and S. Bernath, 1988, *Report of the I.D. team investigation of the Hazel Landslide on the North Fork of the Stillaguamish River* (F.P.A. 19-09420), Department of Natural Resources, Northwest Region, Sedro Woolley, 13 p., plus 2 figures

Booth, D. B., 1989*, Surficial Geologic Map of the Granite Falls 15-Minute Quadrangle, Snohomish County, Washington*, U.S. Geological Survey Miscellaneous Investigations Series, Map I-1852.

Curtis, T., 1988. Mylar overlays showing the historical movements of the Stillaguamish River in the vicinity of the Hazel Slide for the years 1943, 1955, 1965, 1970, and 1978, Department of Natural Resources, Northwest Region, Sedro Woolley.

DNR (Washington Department of Natural Resources), 1996 Hazel Watershed Analysis, Department of Natural Resources, Northwest Region, Sedro Woolley

Miller, D. J., 1995, Coupling GIS with physical models to assess deep-seate landslide hazards, *Environmental & Engineering Geoscience*, vol. 1, 263-276.

Miller, D. J., and J. Sias, 1997. *Environmental Factors Affecting the Hazel Landslide*, Level 2 Mass Wasting Module, Hazel Watershed Analysis, Department of Natural Resources, Northwest Region, Sedro Woolley.

Miller, D. J., and J. Sias, 1998. Deciphering large landslides: linking hydrological, groundwater and slope stability models through GIS, *Hydrological Processes*, vol. 12, 923-941.

Palladino, D. J., and R. B. Peck, 1972, Slope failures in an overconsolidated clay, Seattle, Washington, *Geotechnique*, vol. 22, 563-595.

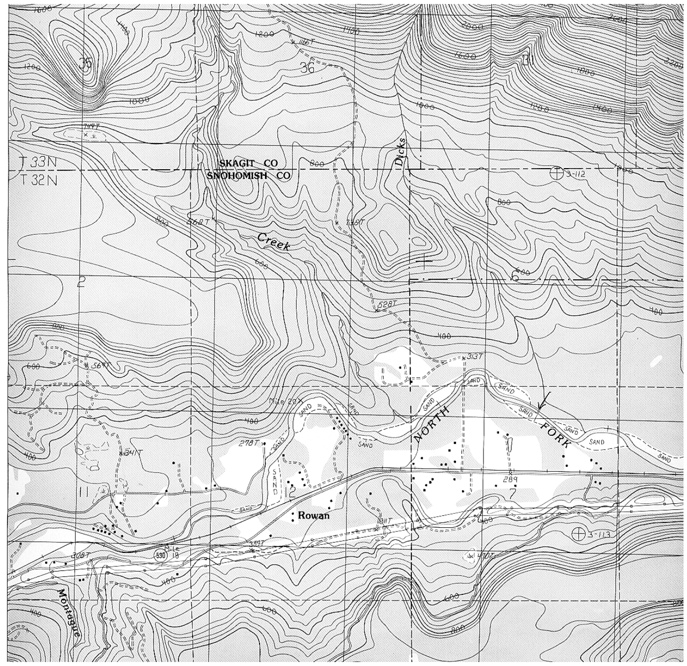
Shannon, W. D., and Associates, 1952, Report on slide on North Fork Stillaguamish River near Hazel, Washington, to the State of Washington Departments of Game and Fisheries, 18pp. plus appendix by H. Coombs, 8 plates.

Tabor, R. W., D. B. Booth, J. A. Vance, and A. B. Ford, 1988. *Geologic Map of the Sauk River 30- by 600-Minute Quadrangle, Washington*, U.S. Geological Survey Open File Report 88-692, 50 p., 1 map sheet.

Thorsen, G. W., 1996. Mass Wasting Module, Level 1 Hazel Watershed Analysis, Department of Natural Resources, Northwest Region, Sedro Woolley.

Figure 1.

40-foot ontours from USGS Mt. Higgins 7.5-minute topographic quadrangle. Reproduced here at a scale of 1:55000.



Shaded relief image of area shown in map above.

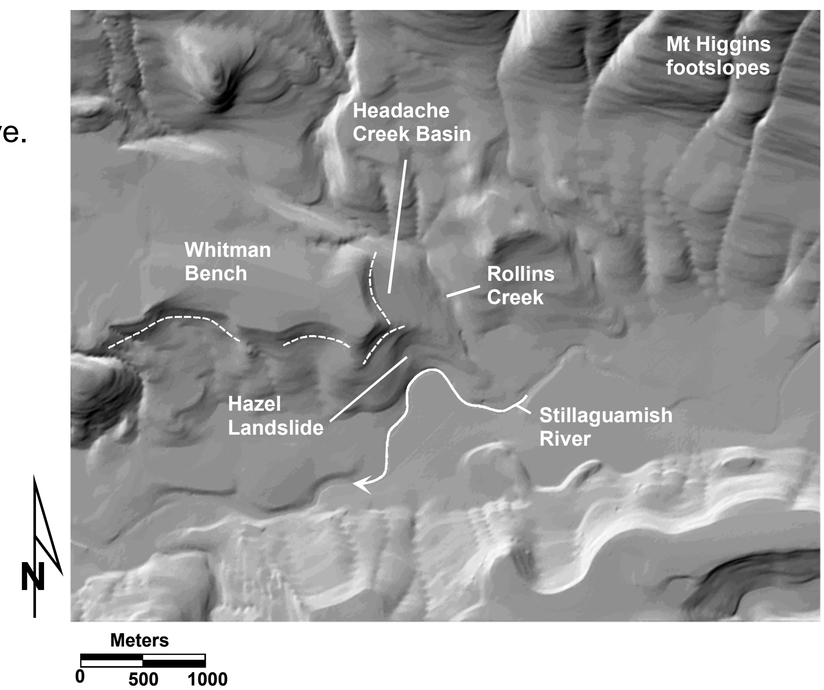


Figure 2. Aerial photograph record for Hazel landslide.
Major events occurred in 1951, 1967, and 1988.

Figure 3. **View of the Hazel landslide in 1967, within weeks after the January event.**

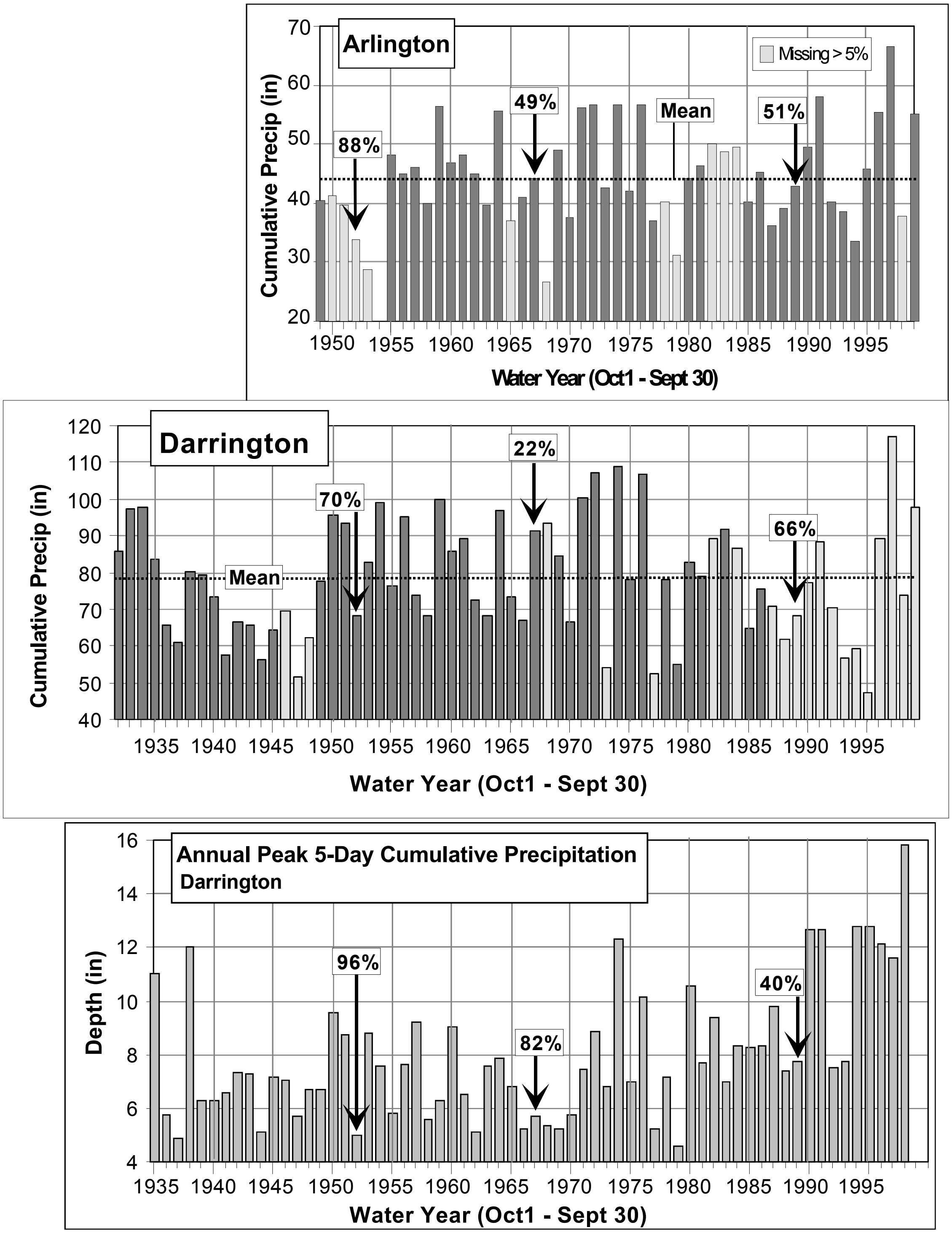


Figure 4. Summary climate data.
Data are taken from weather stations at Arlington and at Darrington. The upper two graphs show cumulative annual precipitation for both gages. Years missing values for more than 5% of the days over the period October to June are indicated with light shading. Timing of the three largest events on Hazel are shown with arrows. Numbers above the arrows indicate proportion of years having greater cumulative precipitation. The lower graph shows peak annual 5-day cumulative totals. This provides a measure of storm intensity that includes antecedent moisture.

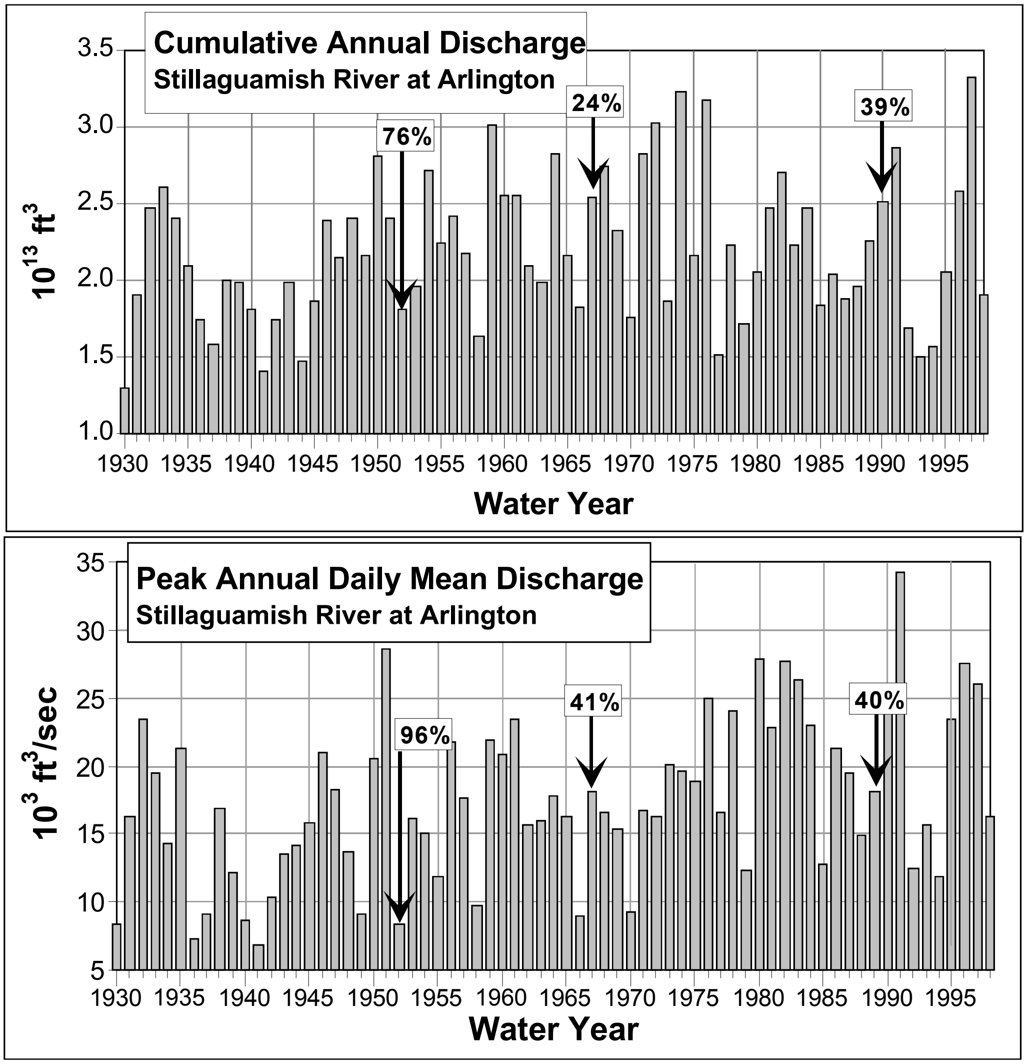
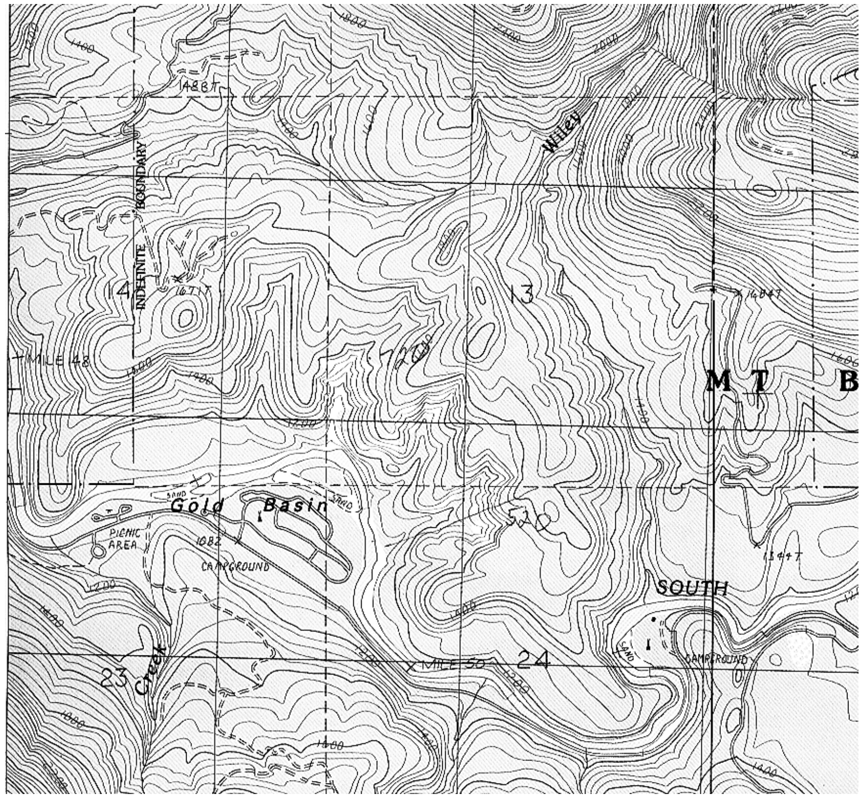
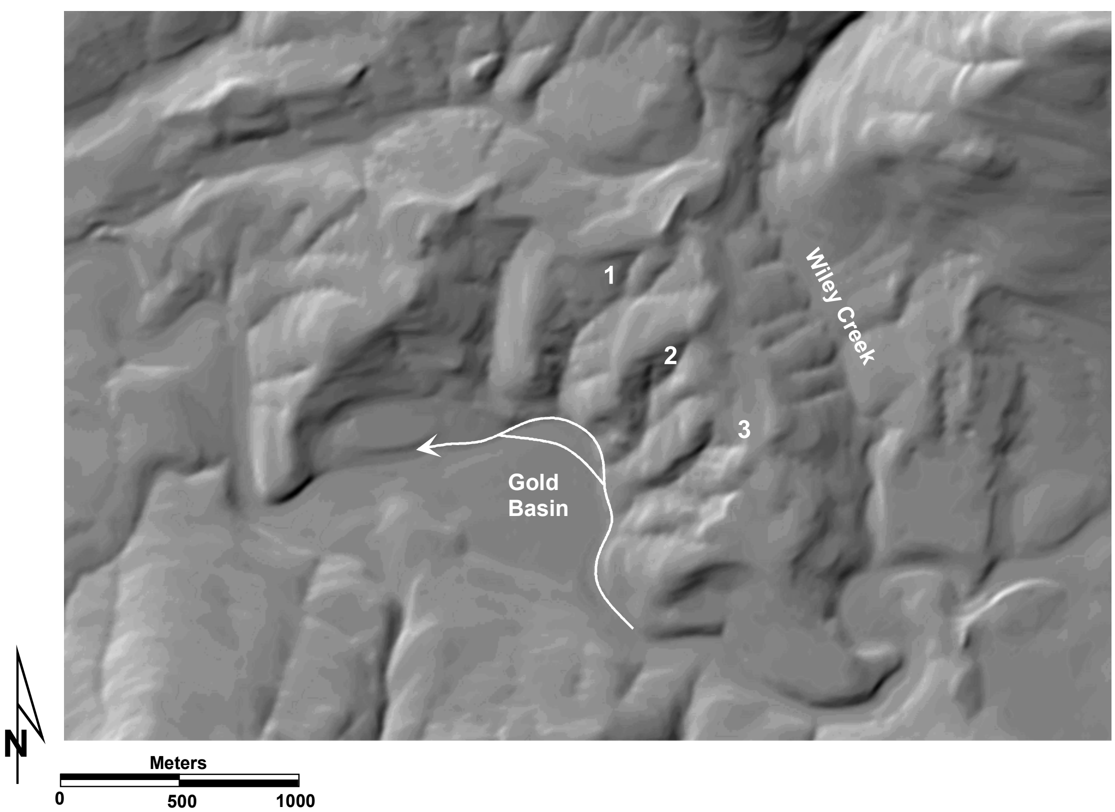


Figure 5. Summary discharge data for the North Fork Stillaguamish.
These data are taken from the gage near Arlington, which has the longest continuous record. The upper graph shows cumulative annual discharge, the lower graph shows annual peaks. Again, the timing of major events at Hazel are indicated with arrows. The numbers above the arrows indicate the proportion of years having larger cumulative or peak values.

Figure 6.
Topography of the Gold Basin landslide. Upper figure shows 40-foot contours from the Mallardy Ridge USGS 7.5-minute quadrangle, here reproduced at a scale of 1:30000.
The lower figure shows a shaded relief image for the area. The three lobes of the Gold Basin landslide are identified with numbers.





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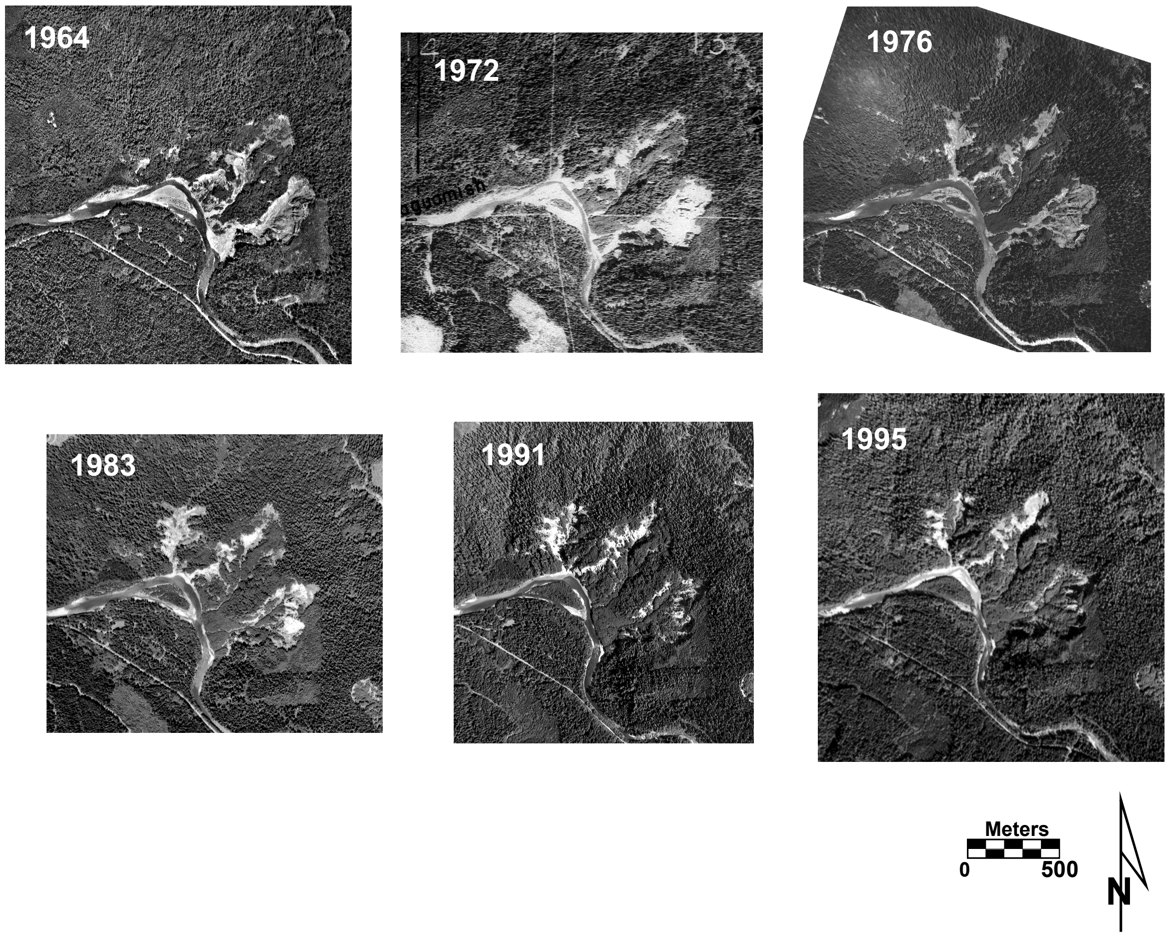

Figure 7. Aerial photograph record for Gold Basin.

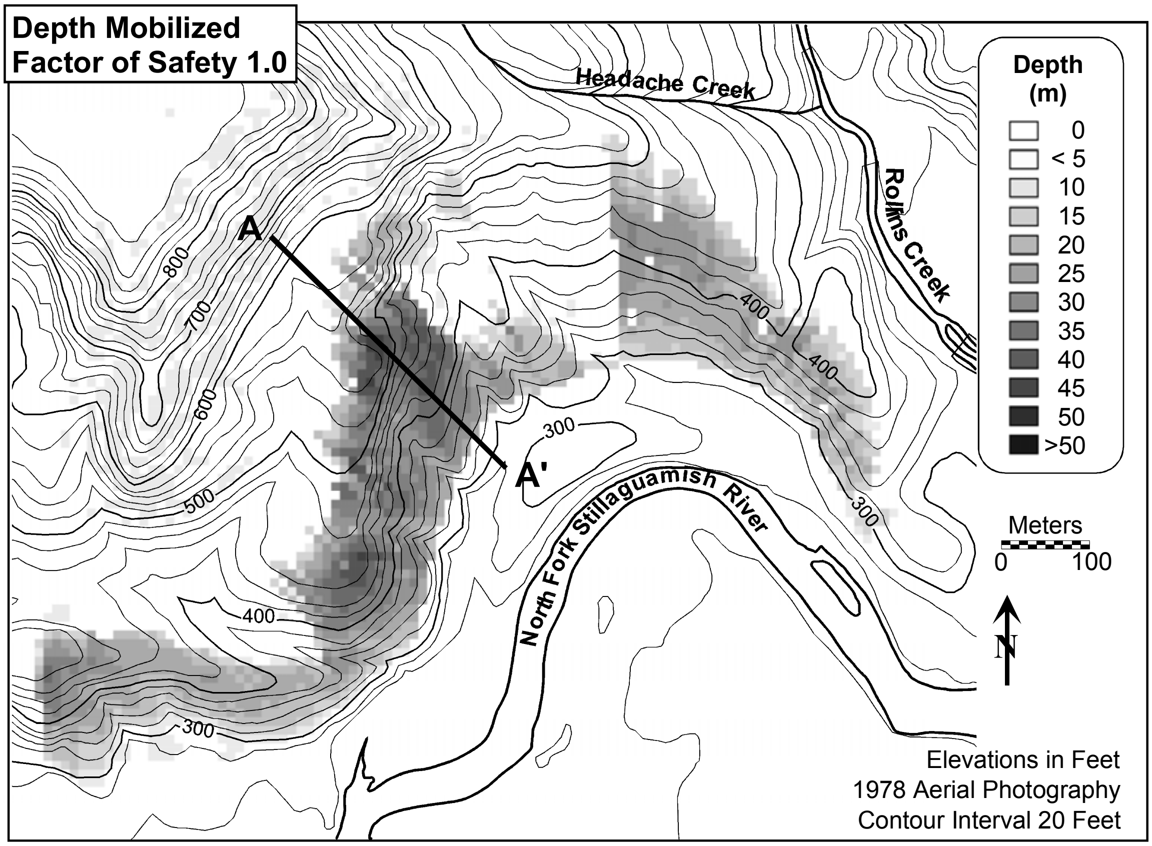

Figure 8. Estimated slip-surface depth.
Using the model described in Miller (1996) and in Miller and Sias (1997), slip-surface geometry is estimated at all points on the landslide for a factor of safety of 1.0. Areas with no shading had factors of safety greater than 1.0. Topography was altered from the contours shown by moving the river to its 1984 location. The total volume potentially mobilized along the transect A-A' is 5425 m^3.

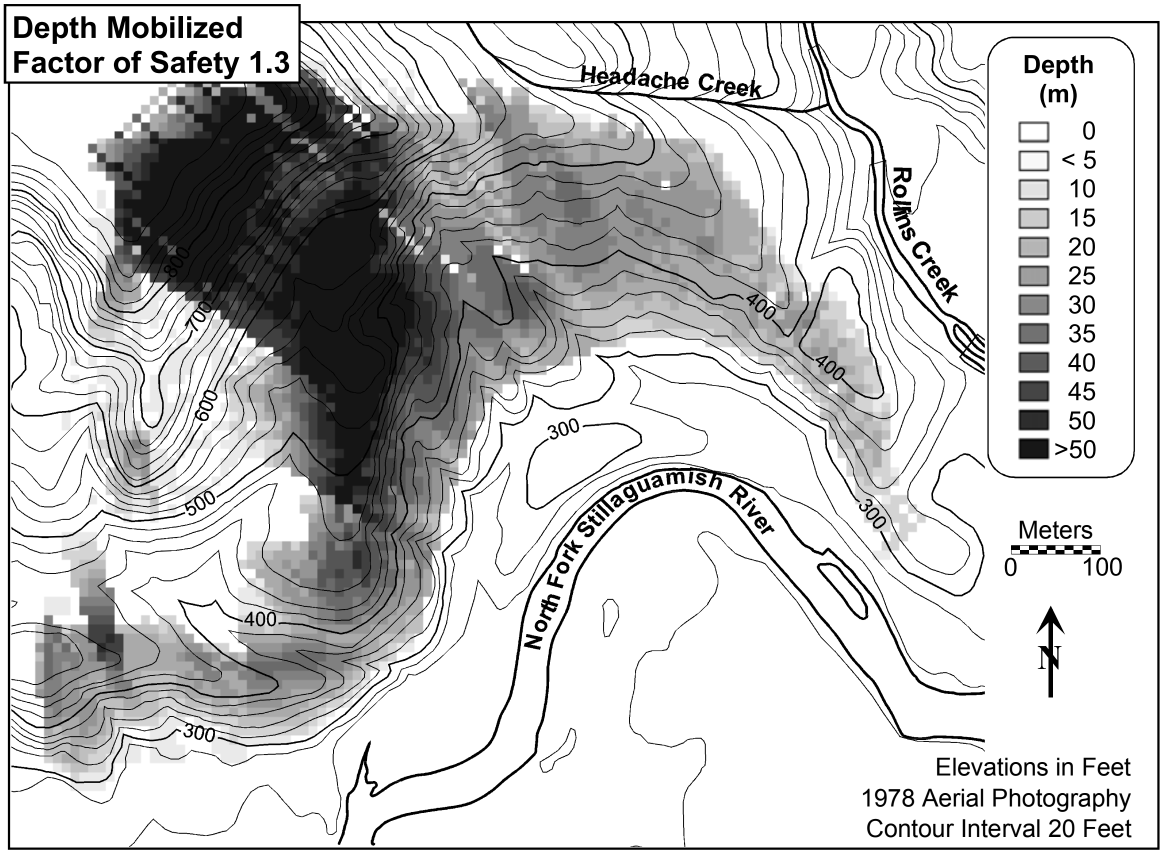

Figure 9. Depth to slip-surface for factor of safety of 1.3.
This figure illustrates the potential effects of further destabilization of the landslide. Progressive failure and unloading of the toe of Whitman bench could initiate movement of a vast volume of material, an order of magnitude greater than that shown in Figure 8.