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LARGE ROCKSLIDES: THEIR CAUSES AND MOVEMENT ON INTERNAL SLIDING PLANES

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ABSTRACT There are two important pre-conditions for the occurrence of large rockslides: first, a long period for rock disintegration on a mountain slope; and second, the long-term persistence of a stable, supporting abutment lower on the slope. Without such temporary support, the increasing volume of the disintegrating material would be prevented by small-scale, downslope mass movements.

Typical consequences of the large rockslides are their movement on internal sliding planes and the development of secondary rockslides. The occurrence, size, and number of the secondary slides mainly depend on the available kinetic energy of the primary rockslide. These findings are based on fieldwork experience on major rockslides in the Alps, the Chilean Andes, and Mount St. Helens, U.S.A.

RÉSUMÉ *Glissements rocheux importants : Leurs causes et leur déplacement sur des plans de glissement internes.* Deux conditions principales doivent être présentes pour que des glissements rocheux importants puissent se produire : une longue période de désintégration des roches sur un versant de montagne, et la présence à long terme d'une butée stable plus bas sur le versant. Sans un tel appui, l'accumulation des produits de désintégration des roches serait empêchée par de petits mouvements de masse plus bas sur le versant.

Les conséquences typiques de glissements rocheux importants sont leur déplacement sur des plans de glissement internes et le déclenchement de glissements rocheux secondaires. L'apparition, l'importance et le nombre de glissements secondaires dépend de l'énergie cinétique disponible du glissement rocheux principal. Ces résultats sont basés sur des travaux de recherche effectués sur les sites de glissements rocheux majeurs dans les Alpes, les Andes chiliennes et le mont Sainte-Hélène aux Etats-Unis.

ZUSAMMENFASSUNG *Massive Felsrutsche: Ursachen und Bewegung an inneren Gleitflächen.* Für die Entstehung von massiven Felsrutschen gibt es zwei wichtige Vorraussetzungen: Erstens, Felszersetzung am Berghang über einen langen Zeitraum, und zweitens, die anhaltende Wirkung einer Barriere hangabwärts. Solch vorläufiger Halt verhindert, daß das Geröll in kleinen Einheiten abrutscht, was eine Ansammlung ausschließen würde.

Typische Folgen massiver Felsrutsche sind Verschiebungen an inneren Gleitflächen und das Auftreten sekundärer Abrutsche. Die Häufigkeit, Größe und Zahl der sekundären Rutsche hängen von der verfügbaren kinetischen Energie des ursprünglichen Felsrutsches ab. Praktische Erfahrung vor Ort an massiven Felsrutschen in den Alpen, den chilenischen Anden und am Mt. St. Helens, USA, bestätigten diese Feststellungen.

INTRODUCTION

The occurrence of very large rockslides is dependent on a series of special conditions; they produce particular effects, some of which are related to their size. This paper provides an analysis of these conditions, processes, and their effects.

According to Heim (1932), a prior condition for the release of a very large rockslide is a long period of time necessary for extensive rock disintegration and weathering on the slope. The longer the time that is available for such disintegration, of course, the larger will be the size of the mass movement when it eventually occurs. Such a process will set off only a single major release, however, if the mass of disintegrating material remains more or less in situ on the slope rather than gliding gradually over time. Thus, the eventual size of a major rockslide will depend on the presence of slope abutments to support the increasing volume of disintegrated material. Both long-term disintegration and partial temporary stability are required. Additional typical features of such major events are the secondary slides. Examples are taken from field studies undertaken by the author in the Alps, Chilean Andes, and Mount St. Helens, U.S.A.

FACTORS FAVORING THE LARGE SIZE OF THE ROCKSLIDES

The relief of high mountain regions is the result of long-term selective erosion and denudation. Nevertheless, these processes do not completely relate to differences in rock stability. This is due primarily to the fact that fluvial erosion, as the dominant process, is linear, regressive, and connected with convergent valley systems. Thus, some unstable areas remain in comparatively high positions in the high mountain relief. Certainly, rocks

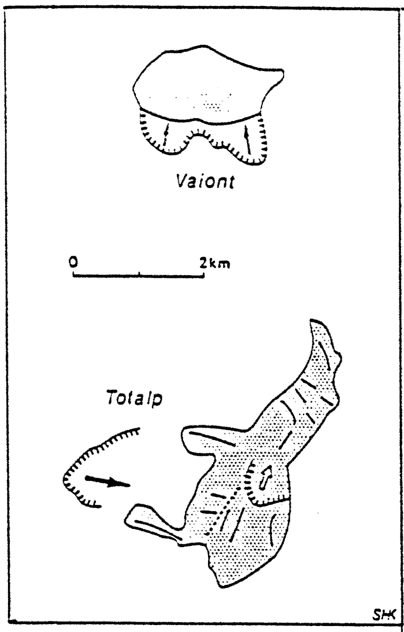


Figure 1. Upper, the Vaiont rockslide, east of the Piave valley in the Italian Alps, with no secondary rockslide.

Lower, the Totalp rockslide, northeast of Davos, Switzerland, with secondary rockslide.

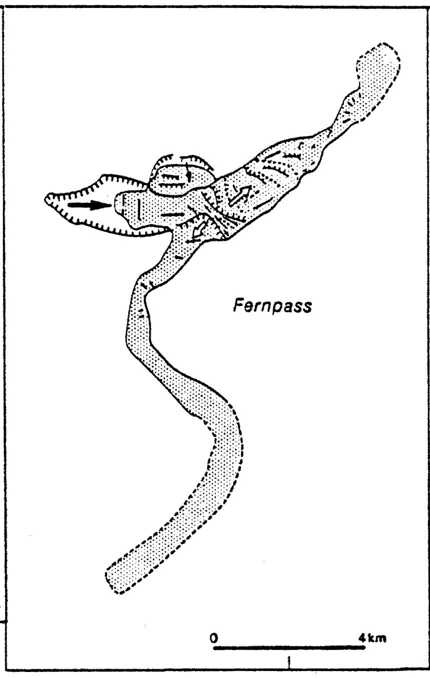


Figure 2. The Fernpass rockslide, Austria.

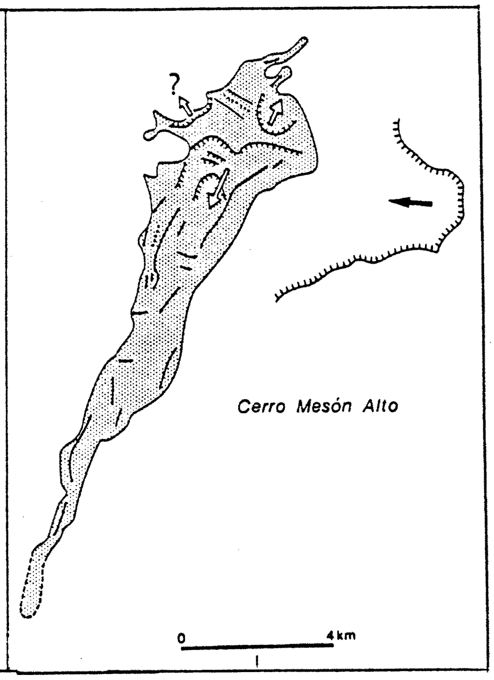
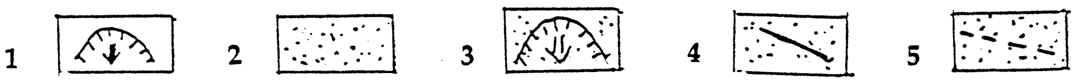


Figure 3. Rockslides from Cerro Mesón Alto, Santiago de Chile.



Key to Figures 1–5:

1. Headscarp of primary rockslide; 2. Area of continuous and discontinuous rockslide debris; 3. Headscarp of secondary rockslide; 4. Marginal, longitudinal, and transverse ridges; 5. Longitudinal and transverse depressions.

of less resistance are subject to processes other than fluvial erosion, such as denudation and especially glacial erosion. Even so, these processes are also strongly influenced by the fluvial forms so that the discrepancy between structure and relief may remain. This may lead to a situation where abutments of stable rock support less stable material. This condition of quasi-equilibrium will be maintained until the abutments themselves are at least partially, if not entirely, worn away. Then rapid mass transfer of the less stable material will be initiated and, if the abutments give way abruptly, such as after being undercut, enormous mass movements may ensue, including rockslides. The greater this discrepancy between the fluvial relief and the structure becomes, therefore, the more important are the abutments, and the greater will be the subsequent mass movements.

In addition to changing from place to place, slope stability also changes through time. This has an influence on rockslide activity. On the slopes of high mountains the progressive disintegration of bedrock by the formation of relaxation joints and by weathering is normally followed by gradual mass transfer. If a considerable time lapse occurs between disintegration and mass transfer, a rockslide or another form of mass movement may be in preparation. Thus, the length of time in preparation (increasing volume of material in a quasi-unstable condition) and the magnidtude of rapid mass movement are closely interrelated. Without the long-term role of the abutment the eventual slide would not attain a great size and potential momentum.

The differences in stability, discussed above, can be seen at the headscarps of some of the rockslides, for instance, at Cerro Mesón Alto (east of Santiago de Chile, 4.5 km^3). In the lower and frontal parts of its headscarp there are granitic rocks. Before the rockslide released, these stable granites supported the less stable andesites, at the upper and rear section, like a dam. There is a similar example at the headscarp of the Totalp rockslide near Davos, Switzerland. Here, comparatively stable crystalline rock (Davoser Dorfberg-Kristallin) supported the less stable serpentine situated behind and above. There are other headscarps where the lithological differences are less distinct. In yet another type of situation, where the material is homogeneous, the differences in stability may be the result of variations in jointing.

Glaciers are very effective agents for initiating rockslides. On the one hand, they oversteepen their valley slopes and remove talus material from their base; on the other, they support the disintegrating mass, thereby preventing downslope gliding. After melting of the ice

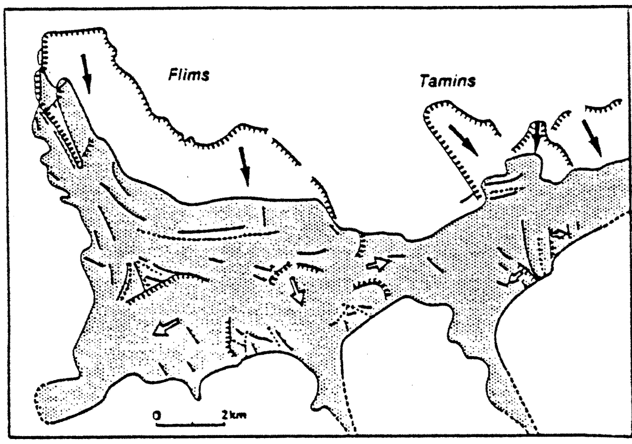


FIGURE 4. Rockslides at Flims and Tamins, Switzerland.

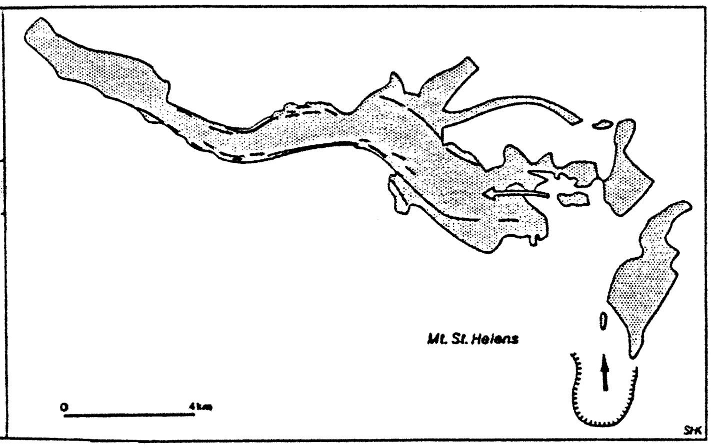


FIGURE 5. Rockslide avalanche on Mount St. Helens, Washington, U.S.A. (after Voight *et al.*, 1981).

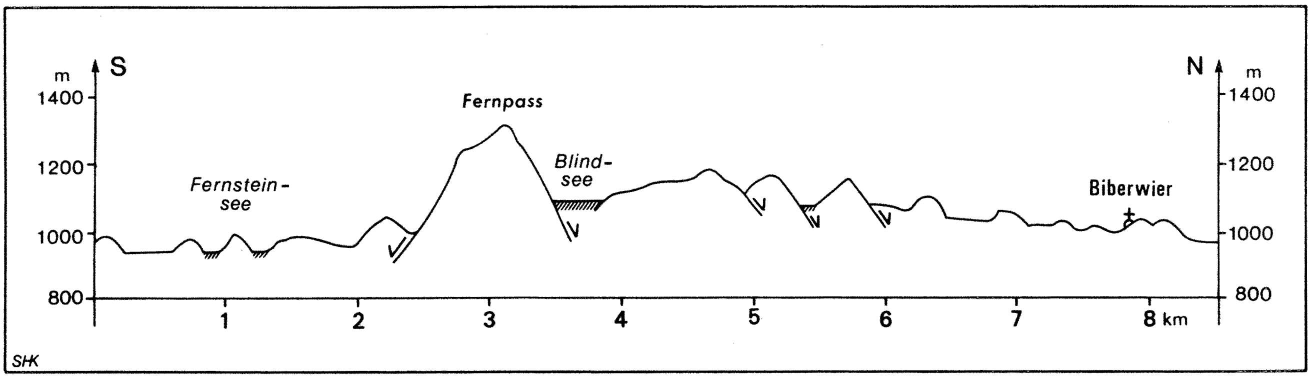


FIGURE 6. Cross-section of the Fernpass area, showing secondary rockslides and headscarps to the north and south; the transverse scarps on the north are the result of tension faults.

and the formation of relaxation joints, the rockslides may release catastrophically and cover the retreating glacier tongue or dead ice, thus eventually producing rockslide moraines. The rockslide in the Río de los Leones valley northeast of Santiago de Chile is a good example of this.

In the Alps the main cause for the descent of large rockslides is certainly glacial over-steepening. Despite this, many of these large mass movements did not release immediately after the initial melting of the ice but released appreciably later in the Holocene. These include: the Tschirgant rockslide (Inn Valley, Tyrol) from which buried wood has provided a radiocarbon age of 2,885 ± 20 yrs. B.P. (Patzelt and Poscher, 1993); the Eibsee rockslide west of Garmisch-Partenkirchen (Bavarian Alps), where debris contains wood of fir trees (Jerz, 1993) with a radiocarbon age of 3,700 yrs. B.P. (mean value of several samples, pers. comm. from H. Jerz, 1994); the Hintersee rockslide (Ramsau, Berchtesgaden, eastern Bavarian Alps), which contains wood radiocarbon dated at 3,520 ± 85 yrs. B.P. (Poschinger and Thom, in press); and the Köfels rockslide (Ötztal, Tyrol) dated at 8,710 ± 150 yrs. B.P. (Heuberger, 1966, and this issue pp. 290–294).

There are several reasons to explain why many rockslides did not release immediately after the retaining ice had melted, including the incomplete segregation of the discontinuity surfaces, or the interlocking of the structures. In some sections the downward trend of the rock masses is not exactly parallel to the neighboring sections. As a result, different rock masses may have a tendency to converge so that a stress field is produced which inhibits downhill movement. Additionally, the jamming and wedging of the different parts of a rock mass may delay downslope movement for a certain time. The delay in response of rockslides after the melting of the ice support could also be due to permafrost. Degradation of permafrost may have determined the time of the release of several of the Holocene rockslides.

A further reason contributing to the initiation of large rockslides is a very dry climate. This can be demonstrated in the driest part of the western slope of the North Chilean Andes near the Tropic of Capricorn. Here the allochthonous Río Grande (North of Salar de Atacama) has eroded a comparatively deep valley; where erosion has dissected the resistant ignimbrites and has reached the subjacent and very unstable San Pedro Formation, many rotational rockslides have occurred on both sides of the valley. These are due to the steep gradient of the slope, which in turn is a consequence of the lack of other slope-forming processes under the predominantly arid conditions.

Aridity is also one of the causes of the great size and number of volcanic rockslides (debris avalanches) in the southern Central Andes: for example, Parinacota 18°10' South, debris area 156 km^2; Socompa 24°25' South, debris area 606 km^2 (Francis and Wells, 1988). In a dry climate, there is less erosion by streams and glaciers and there are fewer glacio-volcanic debris flows (lahars) so that the volcanic cones become steeper and higher. Certainly, the steepness is also due to the predominance of viscous dacite lavas (Francis and Wells, 1988). The greater size and steepness of the cones may lead to a higher incidence of volcano collapse and to particularly large volcanic rockslides (Abele, 1988; Francis and Wells, 1988).

While the number and size of the volcanic rockslides tend to decrease from the dry to the moister parts of the Andes, it is just the opposite with the non-volcanic rockslides. This is due to the increasing depth of the valleys towards the moister areas. There are also more extensive lahars in the moist parts, due to the existence of larger glaciers on the flanks of the volcanoes.

MOVEMENT ON INTERNAL SLIDING PLANES

Nearly all large rockslides start moving downward *en bloc*. This can be seen clearly on the sliding planes of the headscarps where the slides do not leave behind any loose material. If there were a streaming of the whole mass from the start, debris would be deposited along the entire bottom of the headscarp. Subsequent movement is also predominantly *en bloc* because large rock masses have preserved their lithological sequence, for example at Totalp, and even their coherence (e.g., at Vaiont, release 1963, east of the Piave valley, Italian Alps; Flims and Tamins, west of Chur, Switzerland; Köfels; Cerro Mesón Alto). Thus, on the surface of the Köfels slide, for instance, the moraine cover of the pre-rockslide slope has been preserved, even after a long-distance transport on the back of the slide (Heuberger, 1966). The same applies to the northern branch of the Fernpass rockslide (North Tyrolean Calcareous Alps) where the moraine cover has been disrupted by tension faults formed during the slide (see below). This moraine cover had been accumulated by the Inn Glacier on the original slope before the Fernpass rockslide released; thereupon it was transported on the back of the debris to its present position. Thus, in contrast to a former opinion of the author (Abele, 1964), the debris of the Fernpass rockslide was never overridden by the Inn Glacier.

During the gliding of the disintegrated mass, the process of flowing is restricted to thin mobile horizons. The initial movement of the rockslide can take place only if there are primary gliding horizons. Certainly, the mobilizing material of these horizons is soon worn away so that many mass movements slow down before reaching a high speed and covering a long distance (Sackungen). Some of the gliding rock masses, however, attain such a high initial velocity that self-lubrication takes place, contributing to the further acceleration of the slide. The mechanics of self-lubrication have been attributed to the grinding of the rocks by friction (Scheller, 1970), the production of gaseous pore pressure (Habib, 1975), and rock fusion by friction (Erismann *et al.*, 1977).

Very large rockslides can mobilize the water-saturated valley fill into debris flows, which in turn support the further movement of the rockslide material. This happened in the Hinterrhein valley between Bonaduz and Cazis, Switzerland, where sizeable isolated rockslide masses were transported by the debris flow of the "Bonaduzer Schotter" (Pavoni, 1968). The "Bonaduzer Schotter" were mobilized by the Flims rockslide in the Vorderrhein valley. This can be seen at the base of the thick Flims rockslide, where the old valley fill ("Bonaduzer Schotter") is interfingered with rockslide material at Versamer Tobel. Below the confluence of the two rivers, the Vorderrhein and Hinterrhein, the debris flow was dammed by the already-existing debris dam of the Tamins rockslide. That is the reason why it moved up the Hinterrhein valley, thereby incorporating sediments of the Hinterrhein valley and debris of the Tamins rockslide. Valley gravels were also mobilized by the rockslide from the Tschirgant in the upper Inn Valley, Austria (Patzelt and Poscher, 1993) and in the Alm Valley in the Northern Calcareous Alps, Austria. In the Alm Valley the ensuing debris flow transported isolated fragments of rockslide material far beyond the lower fringe of the otherwise continuous rockslide area. There may be many other rockslides where the great distance covered by parts of the debris on the valley bottom can be due to the mobilization of the valley fill (e.g., the Fernpass rockslide, the rockslides of Sierre in the Rhône Valley, Switzerland, and of Kandertal, Switzerland).

Large rockslides move by gliding not only on their base but also on internal planes, where there must also be self-lubrication. It is for this reason that they cover great distances. The geomorphological result of differential movement is the typical rockslide relief of transverse and longitudinal scarps, ridges, and depressions.

If the rockslide collides with an obstacle, such as the slope on the opposite side of the valley, the distal (frontal) parts slow down earlier than the proximal ones. As a result, there is shortening by a sequence of upthrust faults, which leads to transverse ridges.

If, however, the distal parts slow down later than the proximal ones, there is a lengthening by a series of tension faults that bring forth steep transverse scarps with a dip in the direction of the slide. The northern Totalp rockslide (Figure 1) and the northern branch of the Fernpass rockslide (Figures 2 and 6) belong to this type.

The central parts of a rockslide tend to move faster and farther than the lateral sections, thus leaving behind marginal ridges with steep longitudinal shear planes on their inner sides, as at Flims (Figure 4); Eibsee, marginal ridge north of the Hinterbühl depression, at Grainau near Garmisch-Partenkirchen, Bavarian Alps; Cerro Mesón Alto (Figure 3); Mount St. Helens, Washington, U.S.A. (Figure 5). The shear planes are due to the differential *en bloc* movement of the debris.

After losing their lateral sections by the formation of marginal ridges, rockslides encounter more space and tend to diverge. This lateral extension can lead to longitudinal or radial depressions, such as Val Gronda and Val Verena at Flims (Figure 4). The lateral extension and longitudinal depressions are also favored by the increase of lateral space when the rockslide enters a wide valley, and by the splitting effect when it collides with the opposite slope; examples include the depression between Ils Aults and Crest-Aulta, Tamins (Figure 4); Fernpass (Figure 2); and Cerro Mesón Alto (Figure 3).

THE SECONDARY ROCKSLIDES

The collision of a rockslide with an obstacle can lead to the detachment of lateral parts as secondary rockslides which have a direction that differs from the original one. The splitting effect may be the first impulse for this lateral movement. Self-lubrication on the shear planes allows the secondary slides to glide down according to the new configuration and inclination of the relief. The headscarps left behind can be clearly seen in the debris of the primary rockslide.

The occurrence and size of the secondary rockslides depend mainly on the kinetic energy of the primary rockslides. Large rockslides with great vertical displacement are more likely to produce secondary rockslides than smaller ones with little vertical displacement. Thus, the comparatively small Vaiont rockslide (Figure 1) largely maintained its coherence so that even the former relief has been preserved. In contrast, large secondary rockslides were produced by the Totalp rockslide (Figure 1) and by that in the Kandertal, where the volume and vertical displacement are much greater. It is, above all, at the Totalp rockslide that the secondary headscarp is very well developed (Maisch, 1981). There is also a secondary headscarp at the Eibsee debris mass, northeast of the Eibsee (Unterwald).

When rockslide masses are very thick, they tend to develop two or more secondary rockslides. At the Fernpass, very long secondary slides detached themselves on both sides of the primary rockslide debris, where the two secondary headscarps are clearly visible (Figure 2). The primary rockslide from the Cerro Mesón Alto also has two or three secondary scarps on their flanks and a very long secondary rockslide to the south (Figure 3).

The crystalline material of the very thick Köfels rockslide in the Ötztal for the most part retained its coherence so that there are only comparatively small secondary rockslides. Nevertheless, the step-like surface at the northern flank of the debris is due to the northward movement of a secondary rockslide. The steps were formed by tension faults.

The very large Flims rockslide (Figure 4) produced three secondary rockslides, one to the west, one to the east, and a small one to the south, in the direction of the primary slide. With the formation of the very big secondary headscarp in the west, in the basin of Salums, Tuora, Foppas, and Carrera, the longitudinal and radial ridges and depressions of the primary slide at Val Gronda and Val Verena, have lost their southward continuation. In the south, only a small mass of debris could intrude into a narrow valley (Safiental). Therefore, the corresponding area where the material came from is also small, in the basin of Ransun. In spite of shearing displacement and secondary mass movements, the surface of the Flims slide has retained features of the original slope. Thus, the thickest debris mass, at Ault de Val Gronda, lies at the foot of the large western part of the primary headscarp. Further east, the primary headscarp is much smaller and, at its eastern end, the presence of the valley of Bargis caused an even greater reduction of the rockslide mass. This led to the formation of the depression in the debris surface at the foot of the easternmost primary headscarp on the aggradation plain of the Prada, south of Mulin. Certainly, some of the original features of the rockslide surface have been lost by glacial erosion during the advance of the local glaciers from the north (Staub, 1938). It is open to question whether the Rhine Glacier also reached the Flims rockslide area. The crystalline erratic blocks and moraines of this glacier cover some of the southern parts; these moraines could have been transported on the back of the Flims rockslide to their present position. In that case, it is difficult to explain the moraines in the secondary headscarp area north of Salums, Tuora, and Foppas, and in the secondary headscarp area northwest of Ransun. Perhaps the trough-like form of the western secondary headscarp and its moraine cover are due to an advance of the Rhine Glacier which just reached the western part of the Flims rockslide.

If an extremely rapid rockslide collides with the opposite slope, the secondary mass movements include nearly all the material of the primary one. This happened during the Mount St. Helens event in 1980 (Figure 5) where the rockslide avalanche possibly obtained additional mobilization by vapor (Voight *et al.*, 1981). Surprisingly, the comparatively small primary debris mass of the Val Pola rockslide in Valtellina in the Italian Alps which released in 1987, was also extensively distributed. This is probably due to a distinctive disintegration of the rocks prior to the release.

CONCLUSIONS

1. Large rockslides are not only due to long-term disintegration of a slope, but also to long-term abutment of the disintegrating rock masses. Without this support, the disintegrating rocks would be worn down gradually so that no large rockslides could occur. The time of the release of the rockslides is often determined by a change in the topographic, hydrographic, or glacial conditions. Generally speaking, large rockslides are due to discontinuities in space and time.
2. An important reason why large rockslides tend to cover great distances is their movement on internal sliding planes. This differential *en bloc* movement creates the typical rockslide relief with transverse and longitudinal scarps, ridges, and depressions.
3. The occurrence and size of the secondary rockslides depend for the most part on the kinetic energy of the primary rockslides. Small rockslides with little vertical displacement are less likely to produce secondary rockslides than large slides with great vertical displacement. Very thick rockslide masses may develop two or more secondary rockslides.

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