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Talus Slopes

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Introduction

Talus slopes are coarse clastic landforms produced by mass wasting processes. They may be broadly defined as distinctive accumulations of loose, coarse, usually angular rock debris at the foot of steep bare rock slopes. The terms talus (North American) and scree (English) are synonymous and refer to both the landform and its constituent material. Alternative terms include debris slopes (Gardner, 1980) and colluvial fans (Blikra and Nemec, 1998). Talus slopes occur in a wide range of environments but most significantly in those where physical weathering processes dominate and gravitational processes deliver freshly weathered products downslope. The production and/or accumulation of debris must be greater than its subsequent weathering or removal and the accumulations must be of sufficient thickness to develop a characteristic morphology independent of the underlying slope. Simple debris veneers only a few particles thick are termed ‘debris mantled slopes’ (Church et al., 1979). Most studies of talus slopes and talus slope processes have taken place in periglacial or alpine environments. The coarseness of most talus deposits makes them resistant to subsequent erosion: they are often stable, environment. The coarseness of most talus deposits makes them resistant to subsequent erosion: they are often stable, long-lasting elements of the landscape and preserved as relict or inactive forms. Hales and Roering (2005) mapped the distribution of scree slopes in a 80 x 40 km wide transect across the Southern Alps of New Zealand. They estimate that scree only cover 10% of slopes in the wetter areas west of the divide but may mantle up to 70% of the slopes in the drier eastern part, although only 20–25% are presently active.

Talus slopes vary considerably in morphology, composition, and geomorphic setting. Their basic characteristics depend primarily on the morphology and composition of the flanking cliffs, the geomorphic processes involved in their development, and the topography of the cliff foot zone. Although the dominant process is usually assumed to be rock fall, accumulation of talus material and its subsequent reworking may be the result of many different processes acting singly or in combination. This results in a gradational range of forms with several distinctive end members that nevertheless share many common characteristics.

Rock Fall and Rockfall Talus

Rock fall is the free or bounding fall of rock debris down steep slopes. Rock falls vary in size from individual pebbles to catastrophic failures of several million cubic meters (sturzstrom or rock avalanches). Rubble derived from very large single events is usually unsorted, extends much further from the cliff, and may have distinctive morphological features (e.g., ridges and furrows). Normal talus may subsequently develop below the failure surface, covering the upper parts of the feature. Smaller rock falls (< 10⁻¹–10² m³) are the primary process associated with the formation of talus slopes and may be classified into two types (Rapp, 1960). Massive vertical cliffs are dominated by primary rock falls where detachment is followed by direct transfer to the slope below. Primary rock falls are triggered mainly by pressure release, freeze–thaw activity, or weathering on the cliff face. However, on more complex cliff forms, debris may accumulate on irregularities in the cliff (e.g., benches, gullies) and subsequently be dislodged by other rock falls, snow avalanches, surface water flow, or animals. These secondary rock falls have different magnitude–frequency characteristics than primary rock falls. In an interesting discussion of rockfall provision to talus slopes, Krautblatter and Dikau (2007) identify cliff storage as a significant intermediate step between backwearing (destruction of rock from the host cliff) and rockfall deposition (i.e., delivery to the talus), noting that each may have different triggers.

Rockfall Triggers

The triggering mechanisms for rock falls have been inferred from inventory studies that compare the pattern of rock falls with simultaneous observations of temperature and precipitation. Diurnal and, to a lesser extent, seasonal variations in activity are usually studied by direct inventory or observational methods that record the timing of individual events, often some measure of their magnitude, in combination (where available) with meteorological observations from a proximal site. In some cases, they involve periods of continuous observations (e.g., Gardner, 1969, 1980; Luckman, 1976; Rapp, 1960). More generally, they consist of measurement of the accumulation of debris on marked surfaces over time. These have included boulders, plastic sheets (Krautblatter and Moser, 2009; Luckman, 1978, 1988), wooden traps (Prior et al., 1971), annual snow cover (Matsuoka and Sakai, 1999; Rapp, 1960), or netting (Gray, 1972). Decadal-scale inventory data are usually acquired on the basis of the data collected for other purposes, for example, rock falls onto railways (Lan et al., 2010; Rapp, 1960) or onto roads (Bjerrum and Jørstad, 1968; Luckman, 2008; Pack and Boie, 2002; Weiczorek et al., 1992), or through documentary archives. Such studies can yield rockfall volumes, seasonal, and, sometimes, daily records of activity, and again strongly emphasize the seasonal nature of the processes. Recently, several studies have used dendrochronology to address questions of the frequency and spatial distribution of rock falls from sites in Switzerland and elsewhere (e.g., Moya et al., 2010; Schneuwly and Stoffel, 2008a,b; Stoffel, 2006; Stoffel et al., 2010). These tree-ring studies are usually focused on the runout zones or peripheral areas of talus slopes and use scars and anomalous ring characteristics to identify rockfall events, determine rockfall hazards, or evaluate the efficacy of forests as a protection against rock fall. Examination of the position of damage within the annual ring may allow determination of the seasonality of impacts, although
predictably, the majority of impacts occur within the dormant period of growth (Schneuwly and Stoffel, 2008a,b).

At longer timescales, inventory data have been gathered from accumulated volumes on dated surfaces, for example, by lichenometric studies that date individual boulders (Luckman and Fiske, 1995; McCarroll et al., 1998) or the use of lichens to define a dated surface on which new arrivals can be clearly identified (e.g., André, 1997). These studies may allow determination of rates of rockfall activity on decadal or longer timescales and allow inferences concerning the variation of rockfall rates with changing climates, usually the Little Ice Age (e.g., Luckman and Fiske, 1997; McCarroll et al., 1998) or recent global warming (e.g., Gruber et al., 2004).

These inventory and observational data have led to the almost universal acceptance of frost splitting in bedrock and possibly subsequent ice melt as major trigger mechanisms for primary rock falls. This is based on observations of the diurnal or seasonal distribution of rock falls in the mountain and periglacial environments in which rock falls and talus are most frequently studied. In some cases, degradation (thaw) of mountain permafrost with increasing temperatures has also been identified as a potential rockfall trigger in the alpine zone of mountains (Davies et al., 2001; Gruber et al., 2004).

However, detailed observation and modeling of the process is difficult and controversial and has generally been addressed through the context of experimental and theoretical studies on freeze-thaw weathering in periglacial environments. Recent work by Hallet and others (reported extensively by Hales and Roering, 2007, 2009) has suggested that simple volumetric expansion of freezing water within voids or planes of weakness in the rock (the traditional model of freeze-thaw weathering) is not capable of developing the confining pressures necessary to split rocks apart. They suggest that rocks are split by segregated ice growth resulting from migration of water through the rock to the ice lenses at subfreezing temperatures. This process is facilitated by thin interfacial water layers called premelted films and operates at stable temperatures between −3 and −8 °C, well below the fluctuations around 0 °C suggested as significant for conventional freeze-thaw activity. Frequent temperature oscillations across the freezing threshold are therefore unlikely to be a major rockfall trigger under this mechanism. Hales and Roering (2007) present a simple model of the depth and intensity of segregation ice growth corresponding with areas exhibiting maximum rockfall erosion rates and scree formation. This process controversy leads directly into the whole literature on the nature and controls of frost or freezing process in periglacial environments which is discussed elsewhere in this volume. Available data on the timing of rock falls usually remain sufficiently imprecise to determine whether falls are associated with the exertion of pressure during freezing (around 0 °C or at lower temperatures of −3 to −8 °C) or related to the release of frost-separated rocks following subsequent melting of the ice. However, the observed spring and fall timing for rockfall events indicates strong temperature dependence of the rockfall-triggering process.

Although less extensively reported, inventory data also indicate that rock falls are often associated with major precipitation events that destabilize loose materials on the slope. Recently, Krautblatter and Moser (2009) studied accumulation over a 4-year period onto large areas of plastic sheeting (940 m²) placed on talus slopes in the German Alps. They report that over 90% of the rock falls are associated with heavy precipitation events. However, these are mainly secondary rock falls mobilizing previously weathered materials from the cliff zone. Other studies also note a greater frequency of falls associated with precipitation events (e.g., Luckman, 1976) or relationships with the diurnal temperature regime on the cliff face, particularly periods of solar illumination (Gardner, 1980). These observations may also include records of fairly randomly distributed triggers of secondary falls, including root penetration and wedging, animal or human activity on the cliff, in addition to entrainment of material by downslope movement of rock falls, snow avalanches, or water that dislodges loose surface materials.

In addition to these more-conventionally discussed, climate-related, rockfall triggers, two other major trigger mechanisms are discussed in the literature. Several authors (e.g., Bull and Brandon, 1998; Bull et al., 1994; Guzzetti and Reichenbach, 2010; Mills, 1991) have noted patterns of considerable regional rockfall activity triggered by earthquakes. Additionally, several authors have suggested that unloading or pressure release may be a significant cause of higher rock fall, rock avalanche, or landslide activity. This is particularly cited as an explanation for greater rates of rock fall and talus accumulation inferred to have occurred immediately following deglaciation when the downwasting of ice cover removed lateral support from valley sides previously buttressed by ice, leading to a period of accelerated rates of rock fall or rockslide activity (e.g., Hinchliffe and Ballantyne, 1999; Wieczorek and Jäger, 1996).

Rockfall Movement

When rock falls reach the talus, their subsequent downslope movement may be arrested completely or the rock falls may continue to travel downslope in a series of bounces trajectories of diminishing length and height, or by rolling or sliding over the surface (see Dorren, 2003; Dorren et al., 2006). These bouncing boulders may produce characteristic ‘bump holes’ in the surface materials of the slope (Rapp, 1960) and/or considerable damage to vegetation or human infrastructures beyond the talus (see, e.g., Dorren et al., 2006; Schneuwly, 2010).

Movement paths may be very irregular depending on the nature of the impacts and the size and shape of the boulder. Boulders often shatter on their first impact and the independent fragments travel on separate paths. Boulder size and shape are critical. Approximately equidimensional boulders tend to roll or bounce following impact, whereas less compact forms may break up on impact. The downslope movement of dischamped boulders can vary considerably. Discoid boulders landing on edge (i.e., with their long-axis vertical) may generate considerable angular momentum, rotating about their center of gravity, and travel ‘wheel-like’ for long distances downslope. However, should they land on their side (the ‘a–b’ face) or lose their vertical orientation, they will be quickly arrested or slide only short distances over the surface. Downslope trajectories will also vary with the properties of the surface. For example, where discoid boulders land on snow, the boulders may mire...
(landing on edge) or slide depending on their orientation on impact. This erratic behavior is dependent on individual boulder characteristics and the detailed configuration of obstacles in the boulders’ path and makes modeling of rockfall paths and runout distances challenging because of the many potentially random elements involved.

**Rockfall Hazards**

The largest boulders may come to rest some distance below the talus foot, and many talus slopes have a runout zone or a basal fringe of ‘out runners’ (Blikra and Nemec, 1998). As a result of the obvious hazard associated with rockfall, considerable work has been carried out in modeling rockfall activity. Much of this work is summarized by Dorren (2003), Guzzetti and Reichenbach (2010), and Lan et al. (2010). Dorren (2003) divides these models into empirical, process, and GIS-based models (e.g. Marquinez et al., 2003) and they may further be subdivided into two- or three-dimensional models depending on whether the rock falls are allowed to deviate laterally from a straight downslope path. Many talus slopes have a basal runout fringe of boulders (known as the rockfall shadow zone; Evans and Hungr, 1993) extending beyond the lower limit of the talus. Particular attention has been paid to modeling maximum runout distances in this zone because of the obvious hazard to human structures. Early, two-dimensional, empirical models used Heim’s (1932) concept of rockfall fahrboeschung (defined as the angle between the top of the slope and the furthest traveled boulder) to define this rockfall hazard zone. Subsequent work from British Columbia by Evans and Hungr (1993) suggests that the outer limit of the shadow zone is best defined by the intersection between the ground surface and a line projected downslope at an angle of 27.5° from the top of the appropriate talus slope.

Three-dimensional modeling of rockfall trajectories has become increasingly prevalent, and numerical, physics-based, models of rockfall trajectories, draped over digital terrain models of potential downslope pathways and supplemented by GIS analyses, have been used to produce sophisticated models of rockfall trajectories (see Guzzetti and Reichenbach, 2010; Guzzetti et al., 2003 and references therein) that predict or reconstruct rockfall paths and runout distributions. These allow a more accurate prediction of runout distances and may be used to model activity in a variety of settings and at scales ranging from individual sources to regional patterns of activity. For example, Dorren et al. (2006) compare three-dimensional modeling with the results of real-size experiments (boulders ~0.95 m diameter) to evaluate the effectiveness of forest cover as protective devices to mitigate rockfall hazards.

**Rockfall Talus Morphology**

Rockfall talus slopes result from the accumulation of rockfall debris from discrete rockfall events over long periods of time. The plan morphology of talus slopes depends on the form of the cliffs supplying the debris and the morphology of the surface on which the debris accumulates. Relatively simple cliffs, straight in plan view and of uniform or similar geology, tend to produce sheet (straight) rockfall talus slopes lacking significant lateral variation in their characteristics (Figure 1). However, as the cliffs become dissected, rock falls moving downslope are channeled into gullies or basins, and deposition is focused below couloirs, leading to differential accumulation and the eventual development of talus cones. As well as funneling rock falls, these couloirs channel other gravitational transfer processes down the cliff face (e.g., surface stream flow and snow avalanches) that may result in significant modification of the talus below. Therefore, cones developed below dissected cliffs are rarely simple, single-process (rock fall) forms. The upper slopes of the talus may be gullied or reworked by other processes, resulting in more complex landforms (Figure 2).

Many talus slope profiles are characterized by long straight segments at or about 33–35°, often with a strong basal concavity. These slopes have been the focus of considerable field measurement, experimentation, theoretical analyses, and discussion. Initial talus models assumed that these straight slopes were developed by sliding and failure of loose unconsolidated materials (termed dry avalanching or, more recently, grain...

![Figure 1](image1.jpg) A classic straight sheet talus slope developed primarily by rock fall at Small River, British Columbia, Canada. Note the straight slope, slight basal concavity, and ‘run out’ fringe of boulders beyond the talus foot.

![Figure 2](image2.jpg) Coalescing multiprocess talus cones, resting on raised beaches, Templefjorden, Spitzbergen. Note the lower angle, debris-flow-modified cone on the left.
flows) and that the slopes were at the angle of repose of these materials (see Carson, 1977; Statham, 1976). This view was challenged by Statham (1976), who developed a model based on energy expenditure and loss by individual rockfall particles moving over the surface. Extensive reports on measurements of talus slope angles (e.g., Francou and Manté, 1990; Sauchyn, 1986) indicate that measured angles of the straight upper portions of talus slopes are usually between ~32–37° and up to 40°. However, mean talus angles may be as low as 25–30° due to the basal profile concavity. Several authors report differences in mean slope angles as a result of different process combinations (e.g., Church et al., 1979; Sauchyn, 1986). Francou and Manté (1990) propose that talus slopes have a segmented profile: an upper, usually straight slope and a lower section with a basal concavity, separated by a critical angle of ~33–34°. They consider that the upper slope is essentially a transport surface, whereas the lower slope characteristics result from both transport and depositional processes. The length and angle of the basal concavity depend on site characteristics, maturity of talus development, and the depositional processes involved.

Direct observation at many sites suggests that dry avalanches (grain flows) on talus slopes tend to be quite small and are confined to relatively small-sized gravel material on the upper slopes (e.g., Van Steijn et al., 2002). Moreover, examination of many talus slopes indicates that the characteristic coarse openwork surface texture is merely a veneer. The openwork surface acts like a sieve: smaller particles or rock chips landing on the slope are trapped or washed into open voids, which become filled with fines and rock chips at depth. Most active rock falls raise dust clouds from the cliffs and talus, and which become filled with fines and rock chips at depth. Most landing on the slope are trapped or washed into open voids, whereas the lower slope characteristics result from both transport and depositional processes. The length and angle of the basal concavity depend on site characteristics, maturity of talus development, and the depositional processes involved.

### Surface Processes

Talus creep is the downslope movement of individual particles resting on the talus surface. It results from a combination of many processes, including frost action, running water, rockfall impact, needle ice, animal disturbance, and snow avalanches. This process has frequently been monitored using painted stones or lines on the talus in several long-term studies (e.g., Gardner, 1980; Perez, 1993). Rates of movement on upper talus slopes where fines and smaller clasts are mixed typically average 1–10 cm year\(^{-1}\) (greater where more fines are present), but individual stones may move much greater distances as a result of rock fall or avalanche impacts. The movement of large clasts or those on the lower slopes is much less and generally due to burial or dislodgement by direct impact. Several authors have described a process of dry avalanching of materials or grain flows (van Steijn et al., 2002) – though often seen in modeled talus or in gravel piles, this is usually fairly small-scale on natural talus slopes. On upper talus slopes, frost-susceptible fines at or near the surface favor creep through frost action or, in some extreme cases, solifluction. Several authors have reported miniature sorted stripes on the upper portion of talus slopes.

### Talus Materials

Talus deposits consist of angular, irregular rock fragments with a wide range of sizes. The nature of the debris depends to a large extent on the lithology (or lithologies) exposed in the cliff, as joint and bedding characteristics are primary determinants of the shape and size of talus particles. Most talus slopes show a distinctive increase in particle size of surface materials downslope. This is known as ‘fall sorting’, is often logarithmic in form (e.g., Church et al., 1979; Statham, 1976) and is most marked at the base of the slope. Fall sorting is common on rockfall talus slopes, though not universal, and primarily results from two mechanisms. Larger boulders have greater momentum and therefore tend to travel greater distances downslope. Secondly, the frictional resistance (roughness) of the surface over which the boulder slides, rolls, or bounces is defined by the relationship between the size of the moving boulder and the irregularities of the surface (boulders and voids) over which it is passing. Big boulders are only effectively trapped in large ‘holes’ or when they impact other large boulders or obstacles and lose momentum. Therefore, these boulders come to rest in areas of the talus with boulders (holes) of a similar size. The degree of sorting depends on the slope length, cliff height, and the size and shape of dominant particles. It is best developed on slopes with a wide size range of boulders. Locally random effects or differences in particle shapes may result in the absence of or anomalous patterns of sorting. For example, the frictional properties of snow-covered talus are quite different than snow-free talus: large boulders may be trapped in wet snow or slide over the surface and result in local pockets of coarser materials. Processes other than rock fall may modify these sorting patterns (see ‘Modification of Talus Forms’, below).

Several studies of particle shape indicate a tendency for larger particles on the lower slopes to be more equidimensional (e.g., Perez, 1989). Unless they are small fragments, platy or elongate clasts tend to break up with repeated impacts. Also, ‘blocky’ clasts rolling or bouncing downslope can rotate about any axis with similar efficiency, whereas ‘discs’ or ‘rods’ can lose momentum rapidly if their axis of rotation is changed by impact, for example, from rolling to sliding. Talus fabric is also, to some extent, lithology-dependent. Where clasts are elongate or platy, the fabrics tend to dip downslope and plunge less steeply than the talus slope, giving an upslope imbrication (Perez, 1998). Elongate clasts may show a ‘rolling’ fabric (long axis across slope) or a sliding fabric (long axis downslope; Blikra and Nemec, 1998). However, given greater equidimensionality of debris on many lower slopes, fabrics are often, as might be expected, fairly random, unless the talus material is predisposed to an elongated shape by the lithologic characteristics of the material or emplaced by a process favoring sliding; for example, Perez (1989) describes a strong downslope orientation for blocks at Lassen Peak where emplacement is dominated by sliding over a seasonal snow cover.

In recent years, there have been detailed studies of the sedimentology of talus and related deposits (Bertran et al.,
that have developed criteria to differentiate between talus, snow avalanche, debris flow, and other mass-wasting deposits found within talus-like formations. Blikra and Nemec characterize talus as a colluvial deposit consisting of "tongue shaped beds with marked upslope fining, normal grading and mainly openwork texture" that may show a 'rolling' downslope or random fabric depending on the dominant process and the lithologically determined shape of clasts. The deposits of most other processes that influence talus slope development contain a matrix of fines. However, the differences in sedimentological characteristics in these sediments are subtle and difficult to categorize precisely in the absence of surface morphological evidence.

Bedded or 'stratified screes' (grès litées) were first described from periglacial deposits in Northern France (see van Steijn et al., 1995). They are composed of platy or friable calcareous rocks that show dipping beds of coarser openwork gravels alternating with matrix-filled gravels. Studies of these deposits suggest that the openwork beds result from dry grain flows, whereas the matrix-supported deposits are associated with debris flows, sheet flow, or quasiaulvial processes. Recently, Hétu and others (Hétu and Gray, 2000; van Steijn et al., 2002) have described contemporary analogs for these deposits from talus slopes developed from friable, shaley materials in Gaspésie, Québec. Hétu has also described small 'frost coated clastic flows' (FCCFs) that are similar to grain flows, but the matrix-supported deposits are associated with debris flows, sheet flow, or quasiaulvial processes. Recently, Hétu and others (Hétu and Gray, 2000; van Steijn et al., 2002) have described contemporary analogs for these deposits from talus slopes developed from friable, shaley materials in Gaspésie, Québec. Hétu has also described small 'frost coated clastic flows' (FCCFs) that are similar to grain flows, but the matrix-supported deposits are associated with debris flows, sheet flow, or quasiaulvial processes.

Modification of Talus Forms

As talus deposits accumulate, geomorphic processes other than rock fall may rework loose surface materials, modifying talus morphology and surface characteristics. These changes are most frequently seen in periglacial or alpine environments where long-continued modification of talus can lead to the development of distinctive landforms associated with snow avalanches, debris flows, and permafrost creep.

Snow Avalanches

In arctic or alpine areas, snow avalanches running down the cliff face may entrain material and deposit it on the talus below (Rapp, 1960). However, as these snow avalanches move over the talus, they may also erode or incorporate loose surface material, frequently exposing fines at the surface. This material is transported downslope, where it is deposited on coarser material as a scattered cover of precariously positioned smaller boulders or rock fragments, let down from an ablating snow cover. These phenomena are known as avalanche or perched boulders (Jomelli, 1999; Luckman, 1988; Rapp, 1960). The upper, eroded slopes show little size sorting and may be stripped of loose materials, but there is a rapid increase in mean grain size toward the base, where avalanching material is mixed with coarse blocky rockfall debris. Long-continued or intense avalanche activity may carry debris well beyond the limits of the talus, resulting in a loose scatter of coarse debris or the development of an avalanche boulder tongue (Jomelli and Francou, 2000; Luckman, 1978; Rapp, 1959). These 'roadbank tongues' can be recognized by a raised tongue of debris extending beyond the original talus, often with an asymmetric cross-section and a smoothed beveled top, flanked by steep side slopes and a lobate front. Talus slopes modified by avalanches exhibit a pronounced basal concavity, a strong size sorting of surface debris at their lower end and often extend a considerable distance beyond the toe of adjacent, normal rockfall talus across slopes of as little as 8–10° (Figure 3).

Debris Flows

The upper portion of talus slopes contains considerable fines mixed with small rock clasts. Many talus cones are dissected by small gullies that head at the base of the cliff and decrease in size downslope. During heavy, intense rainstorms, water is channeled down the couloirs onto the talus below, mobilizing debris within the gullies on both rock wall and talus, saturating the apex area and generating small hillside debris flows. Single or multiple examples of these features with their characteristic channels, levee margins, and terminal lobes may be generated on the talus, partially covering earlier rockfall-deposited material (Figure 4). Once initiated, these headward gullies will continue to function as pathways for subsequent events, ultimately producing an irregular, often somewhat radial, pattern.
of debris-flow sediments extending downslope across the lower slopes of the talus. Subsequent erosion or dispersal of these materials takes place by later debris-flow events or snow-avalanche activities (Figure 5; Luckman, 1992). These forms have been christened ‘alluvial talus’ and over long periods of time, they may develop into high-angle debris-flow or alluvial cones.

Protalus and Rock Glaciers

Talus slopes are most frequently studied in periglacial environments. At some talus sites, a perennial firn (snow) patch may develop at the base of the slope. Rockfall or avalanche debris landing on this icy surface slides to the base, accumulating as a ridge which has been termed a protalus rampart or nivation ridge.

In many relatively dry, cold alpine areas, thick talus accumulations contain permafrost. The continued buildup of talus and ice in the basal talus may lead to the development of a convex bulge (protalus bulge) in the lower talus. Subsequent deformation and creep of this rock-ice mixture leads to rock-glacier development that transports the coarse talus material away from the base of the slope (Figure 6). Rates of movement are rarely more than a centimeter per year (see Rock Glaciers and Protalus Forms).

Rates of Development

Talus slopes below rock cliffs are a measure of the accumulated removal of debris from these cliffs and therefore have been the focus of many attempts to determine rates of rockwall recession and landscape change in a variety of environments. Attempts at measurement have been carried out using a number of techniques at different timescales. Short-term measurements involve determination of the accumulation on marked surfaces for periods of up to 10–15 years (e.g., Luckman, 1988; Rapp, 1960; Sass and Wollny, 2001). Such estimates are hampered by spatial (the relatively small proportion of the talus sampled) and temporal (only a few years) sampling problems. An extension of these studies has used lichenometry to estimate accumulation rates over longer periods of time (>1–300 years; André, 1997; Luckman and Fiske, 1995; McCarroll et al., 1998). Finally, there have been several attempts to calculate volumes based on well-defined cones, by interpolating the sub-talus bedrock profile from adjacent slopes (André, 1997; Hales and Roering, 2007; Rapp, 1958). In recent years, the potential for error has been reduced by using ground-penetrating radar to estimate the sub-talus surfaces (Sass and Wollny, 2001). The rates determined by all of these measures are invariably fairly low, often <1 cm per millennium but exceptionally up to 1 m. They vary with environment, techniques used, and cliff geology, with little systematic pattern (see summaries in André, 1997; Krautblatter and Dikau, 2007; Sass and Wollny, 2001). In an attempt to examine these controls, André’s studies in Spitzbergen indicate that rates of rockwall recession rarely exceed 1–2 m during the Holocene, except for sites with glacier-related stress relaxation on rock walls, highly frost-susceptible bedrock, or
excessively high cliffs. Krautblatter and Dikau (2007) present an extensive discussion of these estimates and discuss a series of theoretical models of the relationships between rockwall recession (backwearing), intermediate storage on the cliff, and rockfall supply to the talus below.

The low contemporary estimates of debris production and the presence of many inactive talus slopes in formerly glaciated areas have led to the suggestion that much of the talus accumulation in these areas is paraglacial in origin; that is, these features accumulate rapidly during a period of accelerated mass-wasting activity immediately following deglaciation due to the release of confining pressure in the rock walls and exposure to atmospheric conditions (André, 1997; Ballantyne, 2002; Luckman and Fiske, 1995). Mills (1992) has documented talus slopes, 80–100 m thick with well-developed morphology and fall sorting that have formed over an 8-year period in the crater of Mt St. Helens. This indicates that classical talus forms can develop quite rapidly, although in this case, the process is considerably accelerated by abundant freshly shattered bedrock and continued earthquake activity. Nevertheless, the presence of relict or vegetated talus in many areas and relatively low rates of accumulation in some contemporary studies have led to an interesting discussion as to whether the high rates of accumulation reflect a paraglacial (induced by accelerated rockwall failures following glacier recession) or periglacial (severe freeze-thaw activity in a colder postglacial climate) origin (see Ballantyne, 2002).

### Paleoenvironmental Significance

In recent years, some attention has been directed to the paleoenvironmental significance of talus slopes and deposits. Many of these studies involve the examination of gully sections or trenches cut in the upper parts of talus that are presently inactive (Curry and Black, 2003; Hinchliffe, 1999). Alternating layers of openwork clasts interbedded with finer grade deposits, peats, or paleosols are interpreted as intervals of layers of openwork clasts interbedded with finer grade debris-flow/slopewash deposits (see Krautblatter and Moser, 2009), rather than decade-century intervals with a different climatic regime. Contemporary studies of rainfall-triggered debris-flow activity indicate that relatively low-frequency storms may trigger synchronous events over areas of >10 km² (Luckman, 1992; Rapp, 1960), producing similar stratigraphic evidence on different talus slopes. Similarly, although heavily lichen-covered rock fall or avalanche debris that occurs beyond the limits of contemporary activity indicates that there have been larger events in the past, it does not necessarily indicate that there have been changes in the rate or intensity of activity. It may simply indicate that these larger scale events have recurrence intervals that are greater than the effective range of the dating technique used.

Therefore, given that the magnitude-frequency spectrum of processes that contribute debris to talus slopes is poorly defined and it is difficult to determine whether individual deposits represent a single event or multiple events, paleoclimate inferences from these deposits should be made with caution. In many cases, the deposits may represent individual high-magnitude, infrequent events that have little or no relationship to ‘average climate’ conditions.

See also: Permafrost and Periglacial Geomorphology; Rock Glaciers and Protalus Forms; Slope Deposits and Forms.

### References


