Legacies of catastrophic rock slope failures in mountain landscapes

Kenneth Hewitt [a,1], John J. Clague [b,*], John F. Orwin [c,2]

[a] *Department of Geography and Environmental Studies, Wilfrid Laurier University, Waterloo, Ontario N2L 3C5, Canada*
[b] *Centre for Natural Hazard Research, Simon Fraser University, Burnaby, British Columbia V5A 1S6, Canada*
[c] *Department of Geography, University of Otago, PO Box 56, Dunedin, New Zealand*

Abstract

This review examines interpretive issues relating to catastrophic, long-runout landslides in the context of large numbers of recently discovered late Quaternary events. It links relevant research in landslide science, including some novel or hitherto-ignored complexities in the nature and role of these events, to broader concerns of mountain geomorphology. Attention is drawn to mountain ranges known to have large concentrations of events. In particular, discoveries in three regions are singled out; the Karakoram Himalaya, the coastal mountains of northwestern North America, and the Southern Alps of New Zealand. In each region, many new events, or previously unrecognized complexities, have been identified in the past decade or two. Research on the sedimentology and geomorphology of prehistoric, eroded deposits has been critical to identifying rock avalanches, including many that were formerly attributed to other processes. Discoveries of rock avalanches in the ancient stratigraphic record have helped with the field recognition of rock-avalanche materials and in developing facies models of deposits with complex emplacement histories. The stratigraphic record also provides insights into interactions of streaming rock debris with deformable substrates. Such interactions are responsible for "landslide-tectonized" forms and transformation of rock avalanches into debris flows. Of special interest are runout geometries involving the interactions of rock avalanches with topography or substrate materials, and travel over glaciers. Other emerging issues relate to reconstruction of detachment-zone geometries, and slow, deep-seated slope movements that may trigger catastrophic failure. Most previous landslide studies have focused on individual events or general models, whereas the questions addressed here arise from a comparative approach emphasizing common and contrasting features among events in sets and in different regions. The scale and frequency of landslides in the regions of interest mean they have an important role in denudation, regional landform development, watershed evolution, and Quaternary environmental change. A major developmental factor, largely neglected, is persistent disturbance of high mountain fluvial systems by many successive landslides. Damming of streams and subsequent breaching of landslide barriers strongly influence inter-montane sedimentation and denudation, with particular significance in post-, para-, and inter-glacial contexts. Although an individual landslide appears as a "catastrophe" lasting only a minute or two, its legacy can persist as a morphogenetic influence for millennia or tens of millennia through disturbance of other processes. The influence is permanently felt; in effect, multiple events make the event a "normal" one in regions such as the three considered here.
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1. Introduction

Hundreds of rock slope failures larger than one million cubic meters in volume have been identified in the past several decades, mainly in the world's Cenozoic mountain belts. They are particularly common in the Alpine–Himalayan and Inner Asian ranges, and in the mountains of the Circum-Pacific orogenic belt (Voight and Pariseau, 1978). Of particular interest from scientific and hazard perspectives are areas with large numbers of catastrophic, long-runout landslides (Table 1; Abele, 1974; Voight, 1978; Eisbacher, 1979; Whitehouse and Griffiths, 1983; Eisbacher and Clague, 1984; Cruden, 1985; Hewitt, 1988, 1998; Brabb and Harrod, 1989; Savigny and Clague, 1992; Strom, 1998; Hermanns and Strecker, 1999; Weidinger and Ibetsberger, 2000; Evans and DeGraff, 2002; Abdrakhmatov et al., 2004; Blikra et al., 2006).

* Corresponding author. Fax: +1 778 782 4924.
E-mail addresses: khewitt@wlu.ca (K. Hewitt), jclague@sfu.ca (J.J. Clague), jfo@geography.otago.ac.nz (J.F. Orwin).
[1] +1 519 725 1342.
[2] +64 3 479 8706.

Table 1
Preliminary inventory of known, large (>10^6 m^3) catastrophic rock avalanches in the mountain ranges of the world

Number of known events	Mountain region
>100	Alps, Switzerland and Austria (Heim, 1932, Abele, 1974, von Poschinger, 2002)
	Karakoram Himalaya (Hewitt, 2004)
	Caucasus Ranges, Armenia (Karakhany and Baghdassaryan, 2004)
	Andes, Argentina and Chile (Hermanns and Strecker, 1999, Fauque and Tschilinguirian, 2002, Hauser, 2002)
	Southern Alps, New Zealand (Whitehouse, 1983)
51–100	Alaska-Yukon (Voight, 1978)
	China (Li, 2004, Weidinger, 2004)
	Pamir Ranges, Tajikistan (Schneider, 2004, Vinninchenko, 2004)
	Nanga Parbat and adjacent western Himalaya (Shroder, 1993, Hewitt unpublished field surveys)
10–50	Norway (Braathen et al., 2004, Blikra et al., 2006)
	Tien Shan Ranges, Kyrgyz Republic (Abdrakhmatov et al., 2004)
	Alps, Italy (Eisbacher and Clague, 1984)
	Northern Appenines (Abdrakhmatov et al., 2004)
	Hindu Raj-Hindu Kush, Pakistan (Shroder, 1993, Hewitt, 2001)
	Kun Lun, China-Tibet (Fort and Peulvast, 1995)
	Kazakhstan (Abdrakhmatov et al., 2004)
	Southern Tien Shan, Tajikistan (Vinninchenko, 2004)
	Gissar-Alai Range, Tajikistan (Vinninchenko, 2004)
	Nepal Himalaya (Fort and Peulvast, 1995, Weidinger, 2004)
	Taiwan (Evans and DeGraff, 2002)
	Rocky Mountains, Canada (Cruden, 1985, Jackson, 2002)
	Rocky Mountains, U.S. (Brabb and Harrod, 1989)
	Coast Mountains, British Columbia (Eisbacher and Clague, 1984)
	Mackenzie Mountains, NW Canada (Eisbacher, 1979)
	Coastal mountains, Washington, USA (Fahnestock, 1978)
	Sierra Nevada, USA (Wieczorek, 2002)

"Known events" may represent only a fraction of all rock avalanches, with the possible exception of those in Europe, Japan, and the Rocky Mountains.

In this review, we highlight major emerging issues in the study of catastrophic rock slope failures. We emphasize failure, transport, and depositional mechanisms, and the use of the depositional record to interpret the role of these events in the evolution of mountain landscapes.

We draw mainly upon our knowledge of three of the most impressive orogens on Earth, the Karakoram Himalaya of Asia, the high mountains along the coast of northwestern North America (the St. Elias and Coast Mountains), and the Southern Alps of New Zealand (Fig. 1a–c). The Karakoram and the St. Elias Mountains have some of the greatest relief on earth. Slopes commonly rise more than 3000 m, and in places as much as 6000 m, from valley bottoms to adjacent ridge crests and peaks. The Karakoram Himalaya has been described as the highest and steepest terrain on Earth (Miller, 1984), and parts of the St. Elias Mountains equal it in relief and grandeur. At first glance, the Coast Mountains of British Columbia and the Southern Alps of New Zealand are less impressive, as the elevations of Mt. Waddington and Aoraki/Mount Cook, the highest peaks in these ranges, are only 4019 m and 3754 m, respectively. This impression, however, is misleading; local relief in the Southern Alps, for example, exceeds 3000 m where the west side of the range rises abruptly from a narrow, flat coastal plain. This abrupt topographic change demarcates the Alpine Fault, the boundary between the Pacific and Australian plates. Given their relief, it is not surprising that the Karakoram, St. Elias Mountains, and Southern Alps have some of the highest rates of tectonic uplift on Earth (Searle, 1991). The higher watersheds are snowbound most or all of the year, and high

valleys in the Karakoram and St. Elias Mountains contain the largest concentrations of glaciers outside Greenland and Antarctica (Hewitt et al., 1989). Most of the Karakoram and the Southern Alps, and the entire St. Elias and Coast Mountains, were glaciated during the Pleistocene, and valleys in these ranges have been steepened and deepened by ice. These regions have also experienced significant glacier thinning and retreat over the last century, debuttressing the toes of steep rock walls. This glacial legacy, coupled with the ability of some rocks to stand in steep slopes 1000–4000 m high, frequent strong earthquakes, ongoing tectonic deformation, and orographically enhanced precipitation, favor slope failures of unequalled size and impact. Not until the late 1970s and early 1980s, however, were more than a handful of these large landslides recognized in the Southern Alps (Whitehouse, 1983), the St. Elias Mountains (Rampton, 1981), and the Coast Mountains, and not until the 1990s were they identified in the Karakoram (Hewitt, 1999).

To date, 272 late Quaternary rockslide — rock-avalanche events have been identified in the Trans Himalayan Indus valleys of the Karakoram, Hindu Raj, and Nanga Parbat, mostly from the deposits they have left. Hundreds more remain to be discovered. The events include some of the largest landslides on Earth. Many exceed 100×10^6 m^3, and at least six are larger than 1000×10^6 m^3 (Table 2). Vertical displacements, from the top of the detachment zone to the runout limit, are at least 1000 m and in some cases more than 2000 m. Maximum horizontal displacements generally exceed 5 km, in some cases more than 12 km (Hewitt, 1998, 1999, 2001, 2004).

Rock-avalanche deposits in the valleys of the Karakoram are remarkably well preserved, in spite of high rates of tectonic uplift and erosion. The well preserved record relates partly to the semi-arid conditions that characterize the region. It also illustrates how the landslides themselves have controlled landform development in the Karakoram, as described below. However, an apparent contradiction arises between the historic and prehistoric records. Many more prehistoric examples have been found than historic ones, and nearly all of the former were emplaced on ice-free valley floors; barely 10% traveled onto glaciers. Conversely, all but one historic example happened in the glacierized zone. Of the events recorded in the past 200 years, four occurred in the 1980s and all descended onto glaciers. It seems likely that many more Holocene events have occurred in the glacierized zone. The scattered and limited prehistoric rock avalanches there reflect rapid burial of landslide debris on glaciers by snow or dispersal by ice flow and ablation. Rugged terrain and infrequent visits further reduce the chances they will be recognized. Moreover, rock-avalanche deposits that do survive here, as in other glaciated terrain, have often been
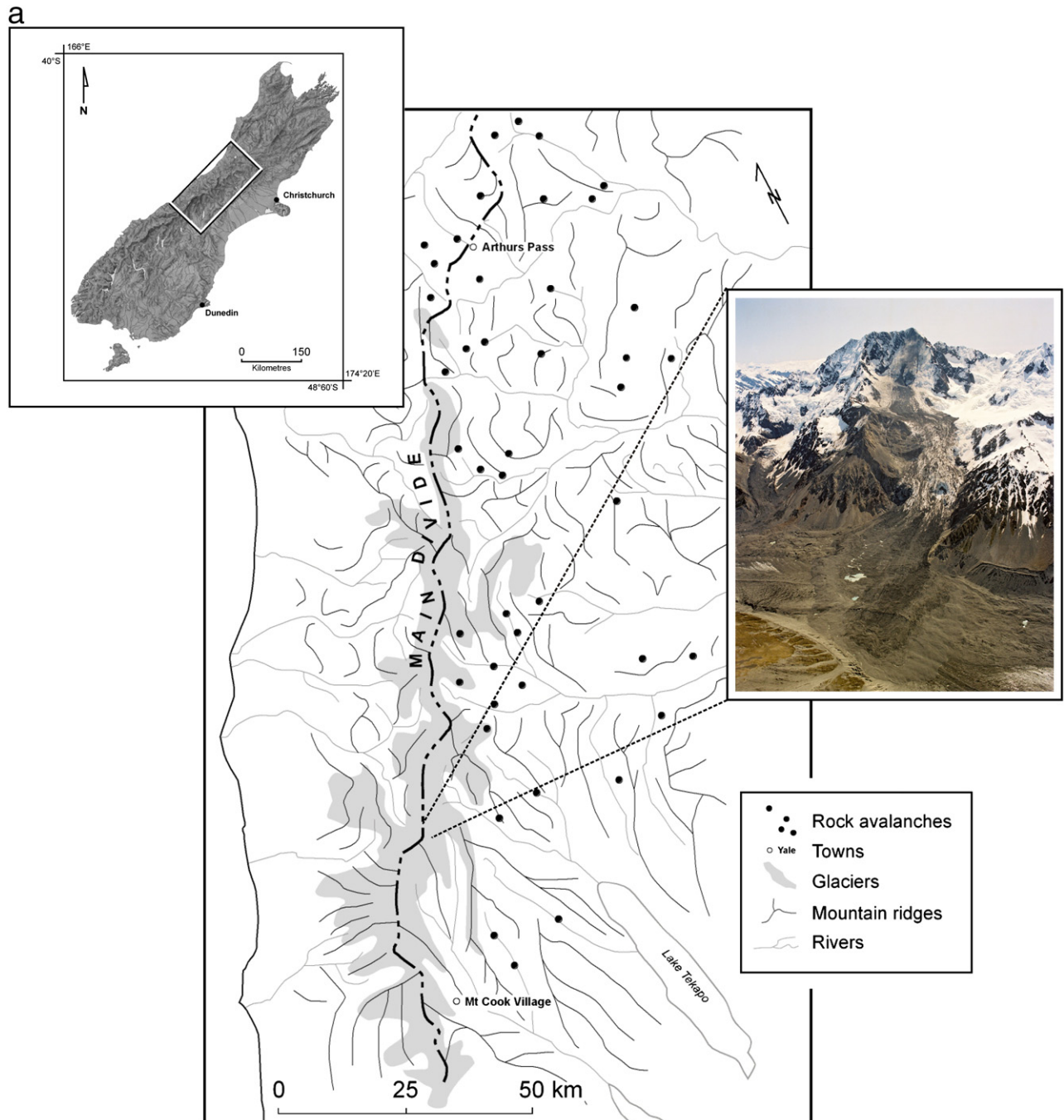

Fig. 1. a–c. Distributions of selected rock avalanches in the Karakoram Himalaya of Asia, the St. Elias Mountains of northwestern North America, and the Southern Alps of New Zealand. Image in Fig. 1b reproduced with permission of Springer Science and Business Media. New Zealand image in Fig. 1c reproduced with permission of GNS Science (photography Lloyd Homer).
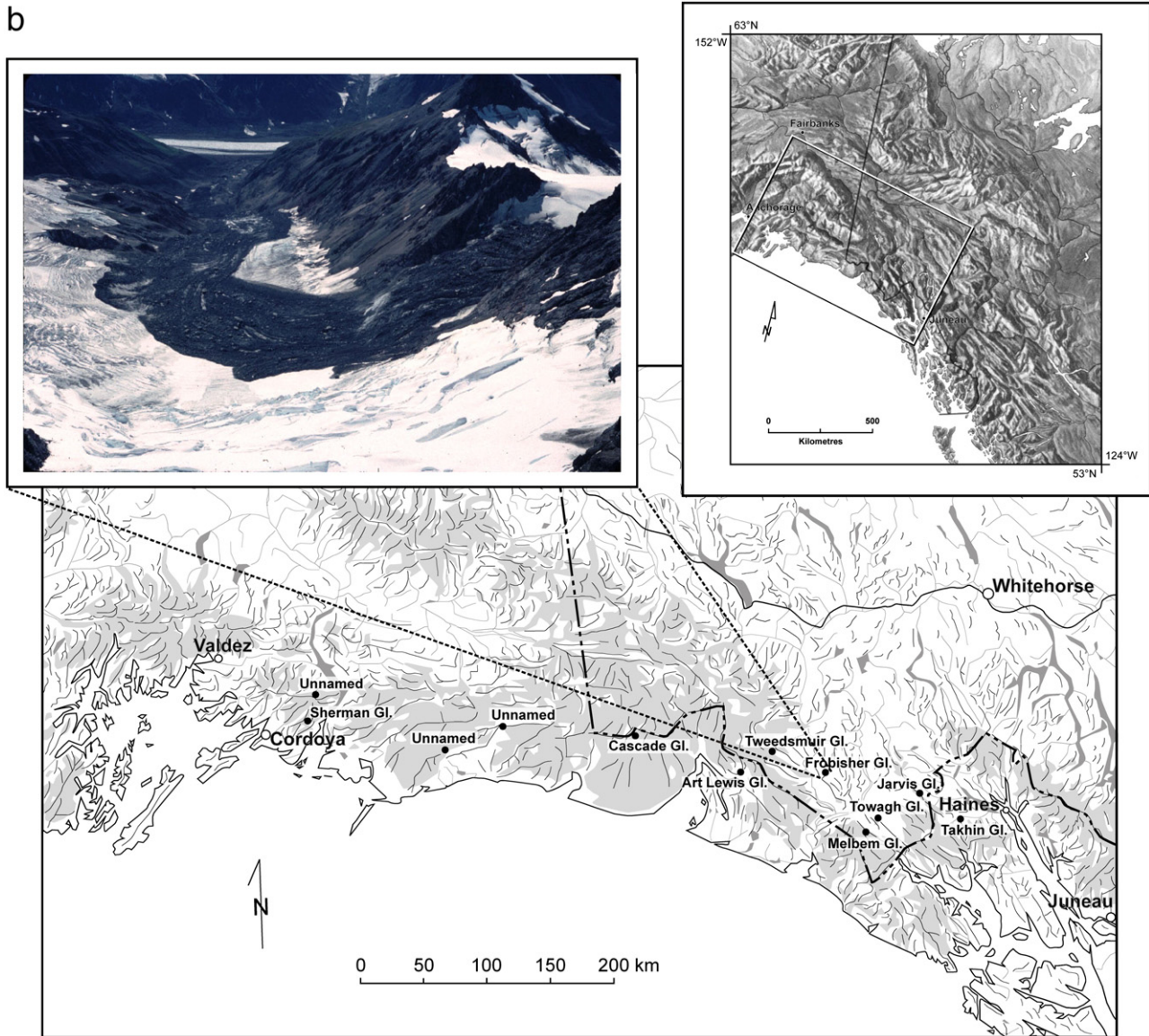
b

Fig. 1 (*continued*).

misidentified as moraines (Heim, 1932; Whitehouse, 1983; Wright, 1998; Hewitt, 1999; Blikra et al., 2006).

Large rock avalanches are also common in the high mountains of North America and the Southern Alps. Many of the historic examples in these ranges have been triggered by earthquakes (McSaveney, 1978; Keefer, 1984; Jibson et al., 2006), although seismic shaking is not a requisite for catastrophic rock slope failure in either area (Cruden and Krahn, 1978; Evans and Clague, 1988; Evans et al., 1989; McSaveney, 2002; Hancox et al., 2005; Geertsema et al., 2006). Most rock avalanches in the St. Elias and Coast Mountains run out onto glaciers and thus are not preserved in the landscape for reasons just indicated. This factor may account for the dearth of old rock-avalanche deposits in these regions. Preservation of rock-avalanche deposits in the Southern Alps is also poor, in part because most are emplaced in relatively narrow valleys where they are rapidly modified or removed by fluvial or glacial processes. In addition, high uplift rates and high precipitation in the Southern Alps favor erosion of

the deposits and their removal from the landscape within decades. Ice cover is much less in the Southern Alps than in the Karakoram and northwest North America. The record of historic rock avalanches on glaciers in New Zealand, however, is considerable and is increasing rapidly, primarily as a result of an improved national seismic monitoring network (McSaveney, 2002, M. McSaveney, personal communication, 2007).

Although the focus of our review is rock avalanches, we do not imply that other types of large landslides are less important. Even greater numbers of translational rockslides, debris avalanches, debris flows, rotational slumps, and mass movements associated with volcanism have been reported in recent years and are important in their own right (Bonnard, 1988; Brabb and Harrod, 1989; Siebert, 2002; Owens and Slaymaker, 2004).

Nevertheless, long-runout, catastrophic rock slope failures involve distinctive processes, landforms, and hazards that have significantly influenced late Quaternary landscape development in scores of mountain ranges. The recent increase in new
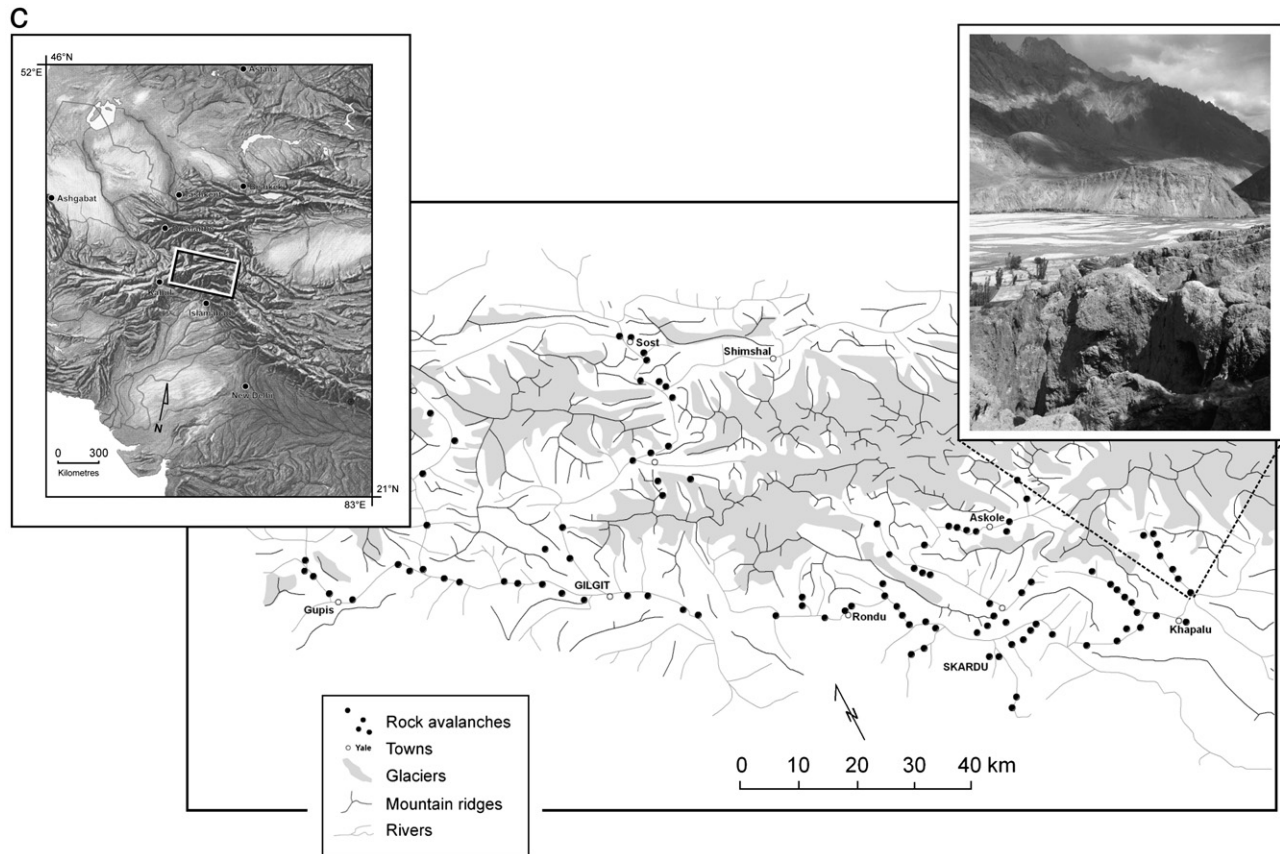
Fig. 1 (*continued*).

discoveries is partly explained by improved access to, and development of, formerly remote, high mountains (Eisbacher and Clague, 1984). However, even more large landslides have been identified in long-settled regions with a century or more of geoscience investigations (Whitehouse and Griffiths, 1983; Hewitt, 2002a; Schuster et al., 2002), and many new discoveries have resulted from the wider availability of high-resolution satellite images (Strom and Abdrakhmatov, 2004). It seems that advances in landslide science are just as important as access. Dozens of deposits formerly attributed to other processes, notably glacial deposition, have been reinterpreted to be the result of catastrophic rock slope failure (Porter and Orombelli 1980; Whitehouse, 1983; Heuberger et al., 1984; Wright, 1998; Hewitt, 1999; Fort, 2000; von Poschinger, 2002; Strom and Abdrakhmatov, 2004).

In general, the abundance of previously unrecognized deposits of large landslides implies a need to reexamine the role these events play in shaping mountain landscapes.

The larger part of the relevant literature on rock avalanches comprises studies of individual events. Most of the rest deals with theoretical and comparative work on either rock-wall stability and landslide-triggering mechanisms, or attempts to model the mobility and runout of rock avalanches (Hungr, 1989; Nicoletti and Sorriso-Valvo, 1991; Legros, 2002; Kilburn, 2004). Here, we consider the importance of these events as Earth surface processes and their contribution to landforms and landscape development. A well developed subfield of landslide

research concerns the formation and failure of large landslide dams (Costa and Schuster, 1987, 1988; Evans and Clague, 1994; Korup, 2002; Abdrakhmatov et al., 2004). The geomorphic consequences of catastrophic rock slope failures, including the landslide dams they may produce, are the direct concerns of a much smaller literature (Clague and Evans, 1987; Hewitt, 1988; Abele, 1997; Shroder, 1998; Hermanns et al., 2001; Weidinger, 2004).

2. Terminology

Landslide terminology in English differs between and within countries, as well as between sub-disciplines such as rock mechanics and geomorphology. Terminological differences partly reflect national perspectives and interests, partly the uneven history of investigations in different countries and orogens, and partly the real diversity of events and regional contexts. Inconsistencies also arise from translation to and from English.

The events we describe and discuss all involve and are initiated by catastrophic failure of bedrock slopes. They are catastrophic in that they occur suddenly, have great size ($>10^6$ m^3), exceptional rates of movement (100–250 km h^{-1}), and are of short duration (minutes). They involve rapid runout of thoroughly broken and crushed rock for distances of several kilometers. The label rock avalanche is widely used and seems appropriate for these landslides (Eisbacher and Clague, 1984; see also below). Rock avalanches are a sub-category of "massive

Table 2
Some large rock avalanches in the Karakoram Himalaya, New Zealand Southern Alps, and western North America

Name	Location (deg)	Volume (10^9 m³)	Vertical drop (m)[a]	Length (km)	Run up (m)	Age (AD or ^{14}C year BP)	References
Karakoram							
Gor-TP	35.30N 74.31E	32	2900	15.5	700	Prehistoric	Unpublished
Nomal	36.03N 75.10E	31	2200	11.0	?	Prehistoric	Hewitt (2001)
Rondu-Mendi A	35.36N 76.25E	23.5	3100	13.2	1100	Prehistoric	Hewitt (1998)
Surmo	35.18N 76.25E	21.5	2580	12.5	600	Prehistoric	Unpublished
Jalipur Complex	35.35N 74.25E	12.3	1600	8.5	540	Prehistoric	Unpublished
Hanichal-JG	35.49N 74.40E	9.2	2050	4.0?	550	Prehistoric	Unpublished
Gor-TP II	35.29N 74.32E	9.0	2100	6.5+	600+	Prehistoric	Unpublished
Basho I	35.18N 76.25E	5.4	2240	9.0+	450+	Prehistoric	Unpublished
Shatial	35.30N 73.33E	4.9	2000	?	450	Prehistoric	Unpublished
Batkor[b]	35.47N 74.29E	4.1	1850	8.6	150	Prehistoric	Hewitt (2001)
Lichar I	35.30N 74.34E	4.0	1850	4/5?	720	Prehistoric	Shroder (1993)
Gol Ghone I[b]	35.17N 75.52E	3.6	1800	7.0	700	Prehistoric	Hewitt (2002b, 2006b)
Telichi Complex B	35.33N 74.33E	3.6	1500	5.5?	350	Prehistoric	Unpublished
Habdas	35.18N 76.25E	2.8	2000	?	540	Prehistoric	Unpublished
Katzarah[b]	35.28N 75.25E	2.1	2300	10.5	750	Prehistoric	Hewitt (1999)
Northwest North America							
Pylon Peak	50.59N 123.52W	0.3	1450	>7	100	7920±100	Friele and Clague, (2004)
Pylon Peak	50.59N 123.52W	>0.2	1450	7.5	–	3930±70	Friele and Clague (2004)
Cheam	49.19N 121.76W	0.17	1600	>5.7	–	4690±80	Orwin et al. (2004)
Hope	49.30N 121.25W	0.05	1100	3.0	–	January 1965	Mathews and McTaggart (1978)
Frank	49.60N 114.24W	0.03	880	2.7	145	April 1903	Cruden and Krahn (1978)
Sherman Glacier	60.53N 145.13W	0.03	600	5.0	–	March 1964	McSaveney (1978)
McGinnis Peak	63.56N 146.27W	0.02	1650	11.0	–	November 2002	Jibson et al. (2006)
Black Rapids Glacier	63.45N 146.18W	0.01	980	4.6	–	November 2002	Jibson et al. (2006)
Mt. Munday	51.33N 125.21W	<0.01	900	4.5	–	1997	Evans and Clague (1988)
Towagh Glacier	59.38N 137.23W	<0.01	880	4.4	–	1979	Unpublished
Tim Williams	56.534N 130.00W	<0.01	880	3.6	30	1956	Evans and Clague (1990)
Southern Alps							
Green Lake	45.76S 167.39E	27	700	2.5	–	12,000–13 000	Hancox and Perrin (1994)
Craigieburn[c]	43.27S 171.59E	0.5	1200	2.7	100	528±96 318±55	Orwin (1998) Whitehouse (1981) Whitehouse (1983)
Mathias River	43.15S 171.14E	0.3	900	1.3	100	1500±390	Whitehouse and Griffiths (1983)
Rangitata River	43.78S 170.76E	0.1	500	3.2	100	3370±880	Whitehouse (1983) Whitehouse and Griffiths (1983)
Jollie River	43.79S 170.24E	0.08	1300	0.8	90	1700±440	Whitehouse (1983) Whitehouse and Griffiths (1983)
Avoca River	43.14S 171.44E	0.06	990	0.9	60	5500±1430	Whitehouse (1983) Whitehouse and Griffiths (1983)
Falling Mountain	42.88S 171.68E	0.06	500	3.2	100	March 1929	Speight (1933) Whitehouse (1983) Whitehouse (1983)
Waimakariri River	43.04S 171.85E	0.04	400	1.8	30	9000±2340	Whitehouse and Griffiths (1983)
Mt. Adams	43.26S 170.53E	0.01	1800	3.0		October 1999	Hancox et al. (2005)
Mt. Cook/Aoraki	43.58S 170.15E	0.01	2720	7.5	70	December 1991	McSaveney (2002)

[a]Vertical distance from top of headscarp to toe of rock avalanche deposit.
[b]Revised from earlier published estimate.
[c]Two events.

rock slope failures," which also include deep-seated, slow-moving landslides in bedrock, some large submarine landslides, and syn-eruptive flank collapses on volcanoes, which we do not address (Evans and DeGraff, 2002).

Researchers who have studied mechanisms of catastrophic, long-runout landslides disagree on terminology and definitions (Hsü, 1975; Hutchinson, 1988; Erismann and Abele, 2002; Collins and Melosh, 2003; McSaveney and Davies, 2007). Many researchers apply nomenclature based on a few fundamental characteristics or laws, while excluding phenomena or events that do not conform to them. However, an interest in all aspects of

specific landslides and how large sets of similar landslides relate to landforms and to the evolution of mountain landscapes requires a more eclectic and inclusive approach. In choosing terms, therefore, we seek to be as consistent as possible, while charting a course through a large and diverse literature.

Another important consideration is that most known landslides, including rock avalanches, are prehistoric and have been reconstructed from eroded or buried deposits. Erosion and burial constrain what can be measured or learned of the original events. Also, many landslide deposits have been incorrectly ascribed to other processes, especially glacial deposition.

Fig. 2. The 1903 Frank Slide in the Canadian Rockies is an example of a rock avalanche with a long runout unimpeded by topography. It emplaced a thin sheet of thoroughly crushed rock 3 km² in area and ca. 2 km from the source at the crest of Turtle Mountain in the distance. The rock avalanche killed an estimated 70 people in the town of Frank at the base of the mountain. (Geological Survey of Canada photo 127435.)

Landslides associated with catastrophic rock slope failures have been labeled rockslide-avalanches (Mudge, 1965), rockfall avalanches (Schuster and Krizek, 1978), rock avalanches, and sturzströms (Heim, 1882; Hsü, 1975, 1978; Hutchinson, 1988; Selby, 1993). The word "avalanche" has been used to emphasise the post-failure phenomena of rapid runout and emplacement of relatively thin sheets of crushed, pulverized, and dry rock. Maximum runout distances are commonly five to ten times the total fall height. The debris generally travels at velocities exceeding 100 km h^{-1}, in some instances more than 250 km h^{-1}. When the velocity of the landslide falls below a threshold, the debris comes to an abrupt halt. Classic examples of rock avalanches include Elm in the Swiss Alps (Heim, 1882), Frank in the Canadian Rocky Mountains (Fig. 2; McConnell and Brock, 1904; Cruden and Krahn, 1978), and the gigantic (>20 km³), prehistoric Saidmarreh event in the Zagros Mountains of Iran (Harrison and Falcon, 1938).

Rock avalanche has been widely used in the literature, even though some researchers contest or reject the term. It provides a graphic impression of what transpires during the landslide, including the creation of a cloud of dust that rises above the streaming debris and emplacement of a sheet of debris over a large area. At the same time, the term does not prejudge the strongly contested issue of how the debris is transported such long distances. Rock avalanches, as defined here, can only arise from sudden, large rock-wall failures and a descent of

some hundreds of meters. Scheidegger (1973) suggested a minimum volume of rock of 0.5×10^6 m³ to develop the streaming behavior characteristic of rock avalanches. Davies and McSaveney (2002) concluded that the transition can occur at volumes as small as 0.05×10^6 m³, whereas others have argued that full development as sturzström requires at least 10×10^6 m³ (Hsü, 1978). Most of the examples discussed here are much larger, exceeding 30×10^6 m³ and in some cases over 1000×10^6 m³.

Rock avalanches are derived from bedrock that was more-or-less in place and more-or-less intact at the moment of failure. Precursory gravitational creep and fracturing are common, but prior weathering and break up of the rock appear to lessen the likelihood of catastrophic failure and runout of the debris. Rather, most failures occur in relatively massive, hard rocks exposed in steep walls of relatively young valleys. Many rock avalanches are derived from the faulted faces of tectonically active young mountains such as the Zagros Range in Iran and the Basin and Range terrain of western North America. Other important contexts are fiord lands, presently glacierized high mountains, the flanks of active and dormant volcanoes, and deeply incised mountain valleys, such as the Himalayan gorges of the Indus and Brahamaputra Rivers.

Of equal interest for mountain geomorphology is the extent to which rugged terrain and deformable substrates may affect the runout process, emplacement forms, and composition of the

deposits. Runout is long, measured in kilometers , but may vary considerably for any given volume and height of fall; so may the extent and character of the break up and crushing of the bedrock material and the emplacement morphology. Moreover, conditions in the runout zone may alter the runout. Large quantities of moisture or sediment may become entrained, changing the distance the debris travels and the morphology and composition of the deposit (see below).

3. Rock avalanche deposits

Most rock avalanches have been identified long after the events themselves, and mainly from their deposits. The landforms associated with these deposits comprise important landscape elements. Their morphology and composition are important for reconstructing events and differentiating them from the many other coarse, unsorted or poorly sorted materials found in mountain regions.
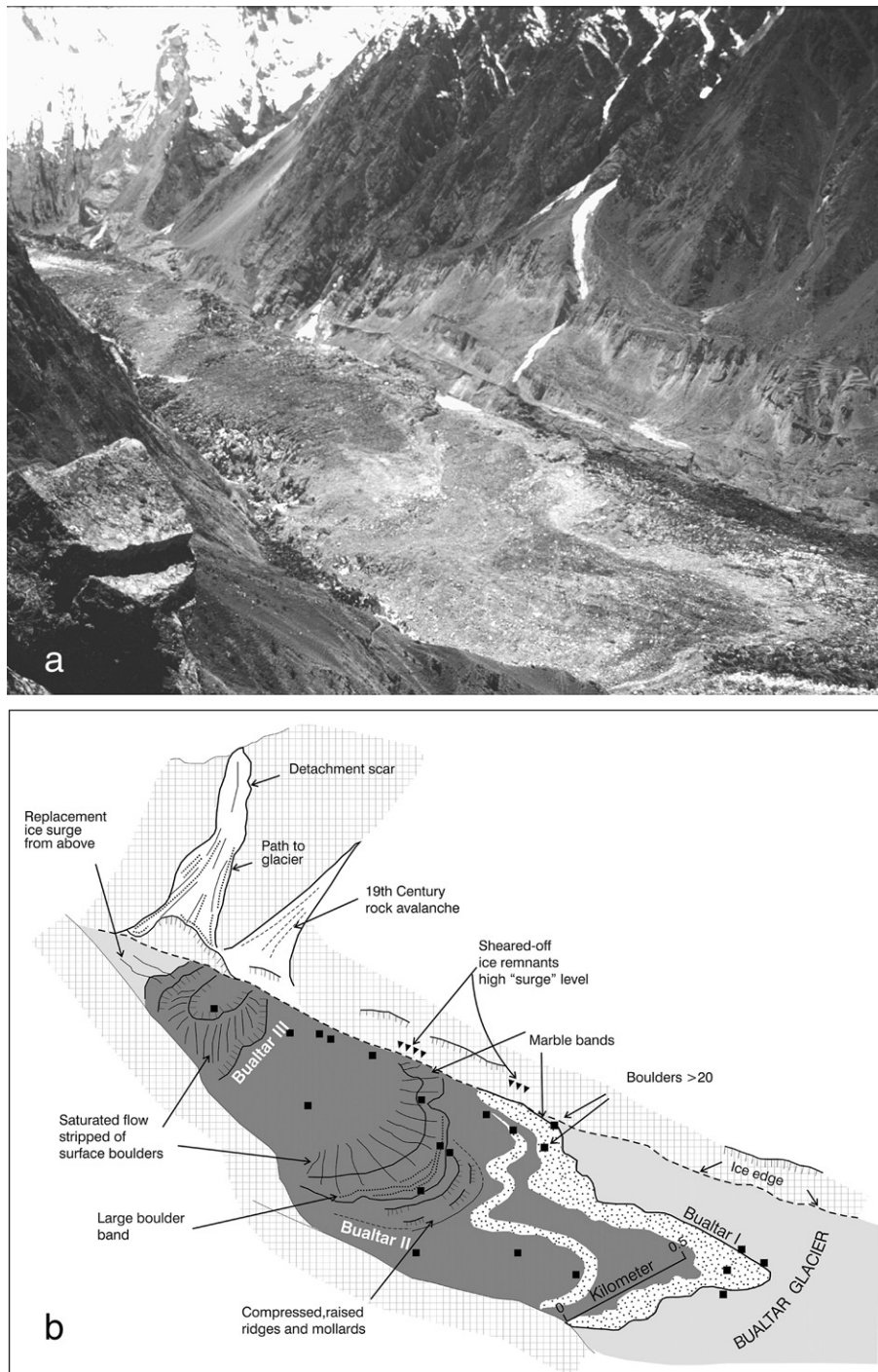

Fig. 3. Remnant stratigraphy preserved in the 1986 Bualtar Glacier rock-avalanche deposit, Karakoram Himalaya. Different lithologies in the source area are preserved in the debris sheet on the glacier. (a) Photograph of debris tongue on Bualtar Glacier. (b) Sketch map of the debris sheet showing lithologic banding, including distinctive white marble bands.

The deposit of a rock avalanche is the product of brittle fracture, crushing, and pulverization of bedrock. In most cases, the fresh surface of the deposit consists of an openwork cover of angular blocks, but these blocks generally overlie debris with a dense matrix of pulverized material (Yarnold and Lombard, 1989; Dunning, 2004). The proportion of matrix is highly variable, both within a rock-avalanche deposit and among different ones, even in the same area. However, sand generally exceeds silt, and, in contrast to many tills, clay is minor. Large quantities of dust are spread over surrounding areas. Debris transport rarely lasts more than 2 or 3 min, but, within this short time, much of the initial failed rock mass is converted to granule-, sand-, and silt-sized material (Blair, 1999; Hewitt, 1999, his Fig. 3; McSaveney, 2002; McSaveney and Davies, 2007). Rock avalanches can, therefore, make substantial contributions to denudation and to comminution of rock.

Sudden collapse of competent masses of hard bedrock, together with the transport mechanism, gives rise to a number of important diagnostic features. In any given sample or at any site, the lithology of the debris is the same, from the largest blocks down to the smallest lithic grains; at smaller sizes, mineral grains occur in proportions similar to those of the parent rock. If the failure involves two or more rock types, the lithologies maintain their identity as uniform bands (Fig. 3), commonly arranged sequentially outward from the source and referred to as "remnant stratigraphy" (Heim, 1932; Hadley, 1964; Shreve, 1968; Hewitt, 1988; Abbot et al., 2002). Minor beds and veins, even textural characteristics of the original rock mass, are preserved as the rock is crushed to powder (Fig. 4). Another important characteristic is that clasts of all sizes are typically angular or very angular. There is little indication of grinding and no indication of polishing, as happens in mass movements with turbulence and particle transport. Rather, the principal comminution process is brittle fracture. Particles of relatively softer minerals may exhibit some snubbing of corners and a few clasts may be scratched, but these features typically are rare.

The shock of the sudden deceleration and cessation of movement converts the main rock-avalanche body into a highly compacted mass. This compactness is one reason why rock-avalanche deposits constitute stable landslide dams.

Many researchers who have studied rock avalanches argue that something other than the comminuted rock mass itself is needed to explain the frictional conditions that allow the rapid and long runout of broken solids. A number of them invoke an intergranular medium, most commonly water in liquid or gaseous form, to account for the extraordinary travel distances of rock avalanches. Formerly, the occurrence of rock avalanches on Mars was seen as a strong argument against these theories, but new evidence makes it far less certain that no moisture or frozen carbon dioxide was involved in their emplacement (Quantin et al., 2004). Yet, whatever the mechanical necessity or wisdom of invoking a fluid, most rock-avalanche deposits show little or no evidence of one. If there were a liquid or gas phase during movement, it must have disappeared by the time the mass halted. There is nothing but crushed bedrock materials in the main body of rock-avalanche deposits and, in most cases, little void space to accommodate fluids. Two hypotheses that do not appear to suffer
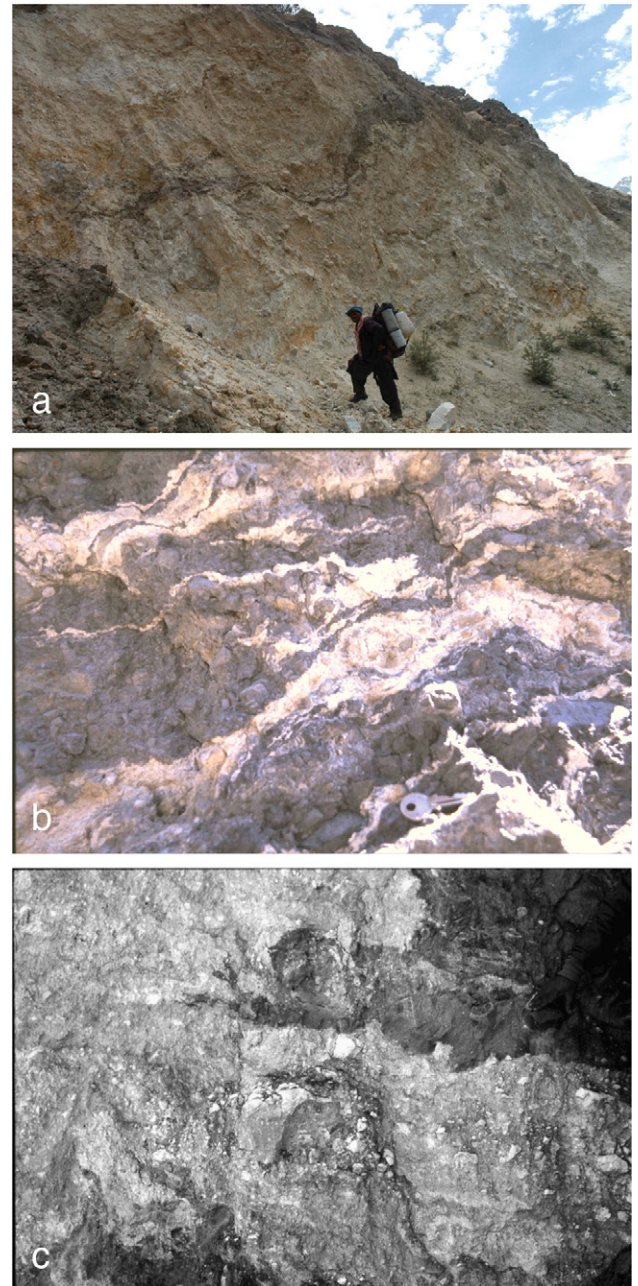

Fig. 4. Lithological and mineralogical differences of the original rock mass may be recognizable in rock-avalanche debris, although distorted by transport and emplacement. (a) Section of the Biaho-Lungma-Hurlang rock avalanche in the upper Braldu River valley, central Karakoram. This exposure is about 10–15 m below the top of the debris lobe near its distal rim and 4 km from the source. (b) Section of the Chaprot-Chalt rock avalanche, Hunza Basin, Karakoram Himalaya. The exposure is about 15 m below the surface and approximately 2.5 km from source; movement from left to right. Source rocks are, for (a) and (b), relatively soft limestone and volcanoclastic sediments of the Yasin Group, Kohistan-Ladakh terrane of Searle (1991). (c) Crushed rock at a depth of 120 m in the Tsok-Dumordo rock avalanche, Panmah Basin, Karakoram Himalaya, about 3.5 km from the source. The darker substrate material, probably glacial sediment, was intruded into the landslide mass. In each of the three cases, the lithological units have retained their identity, although they have been distorted by shear.

from these limitations are the idea of "acoustic fluidization" (Collins and Melosh, 2003) and "dynamic fragmentation" (Davies and McSaveney, 2002). However, neither hypothesis has yet received widespread testing or support.

Other researchers have invoked a low-friction or buoyant layer at the base of the rock avalanche to explain its high mobility, in effect treating it as a slide rather than a flow. Their models involve a critical basal layer of moisture (Abele, 1997), wet sediments (Legros, 2002), trapped air (Shreve, 1966, 1968), or vaporized pore water (Goguel, 1978). However, while snow, moisture, or wet sediments likely contributes to the mobility of some rock avalanches, it has yet to be shown which one or whether any is necessary to the great travel distances of many of them. This conclusion applies especially to rock avalanches that move through narrow canyons, climbing hundreds of meters up steep slopes, and those that remain some tens to hundreds of meters thick, yet deform and fragment at high speeds according to the geometry of their own mass. Examples include the Flims and Köfels events in the European Alps, Usoy in Tajikistan, and Haldi, Gol-Ghone, Nomal, and Rondu-Mendi in the Karakoram Himalaya. Even if there were a component of passive translation on a basal fluid cushion in these landslides, most of the strain and the important dynamic developments took place within, not at the base of, the material. Conversely, when streaming rock debris encounters or entrains snow, ice, water, or deformable sediment, the character of the mass movement changes, resulting in some of the complex or compound events discussed below.

From geologic and geomorphic perspectives, the search for a medium or mechanism of mobility beyond that provided by the rock material itself seems to ignore much of what we encounter in the field. It has diverted attention from issues such as diverse and complicated emplacement morphologies that are empha-sized here. In most cases, the form that is invoked in the various rock-avalanche models mentioned above is atypical (see below). Meanwhile, other questions surrounding rock-avalanche genesis, topographical context, and emplacement and post-emplacement history have emerged as critical to their behavior and their role in landform development and mountain denudation.

Equally important, these events disrupt other geomorphic systems, notably by emplacing huge cross-valley barriers and, in some cases, long-lived natural dams (Costa and Schuster, 1988; Clague and Evans, 1994a; Abdrakhmatov et al., 2004). Their large detachment scars, irregular blocky deposits, and constructional landforms can persist and affect landform devel-opment for millennia or longer.

4. New directions and problems

New discoveries, and reexamination of some well known examples in light of these discoveries, have led to a sense of novel or hitherto-ignored complexities in the nature and legacy of catastrophic rock slope failures. The complexities relate to:

1. runout geometries associated with interactions of streaming debris with topography;

2. interactions with, and incorporation of, substrate materials;
3. interactions with glaciers;
4. reconstruction of the geometry of failed bedrock masses;
5. relations between sackung and catastrophic slope failure; and
6. precursory and triggering mechanisms.

It may seem that this list starts from the wrong end. However, some of the key processes happen well below the surface of the streaming debris, or too quickly, or are obscured by dust clouds. Also, precursory phenomena have rarely been recognized or monitored.

Most detachment areas in rugged mountains are difficult or too dangerous to access. As a result, most events are known only from the more-or-less ancient deposits they produce, and most published studies begin with the end product — the deposit — and work backward to reconstruct the nature or roles of the other phenomena. Rock avalanches that have occurred in recent time, and have been witnessed, have contributed greatly to our understanding of the phenomena (Hsü, 1978), but they offer limited perspectives on the variety of events now known to constitute the longer term record.

Unlike the deposits of recent rock avalanches, older ones may be dissected and segmented by erosion. Erosion may create problems of recognition, but it also exposes the interior of the deposits and may reveal much about flow and emplacement processes. In some cases, the geometry of the initial failed rock mass has profoundly affected the form of the deposit, whereas in others it seems to have had no discernable effect. In this respect, studies of ancient rock avalanches can make a substantial contribution to our understanding of more recent events (Krieger, 1977; Yarnold and Lombard, 1989; Abbot et al., 2002).

Some of the important complexities emphasized here involve blocking and stalling of runout, which prevent full development of the rock avalanche. These processes may, however, pre-serve features that record intermediate phases that many or all rock avalanches go through, but that disappear with fuller development.

In broader terms, we also need to address important emerg-ing issues and innovations relating to:

7. interactions of rock avalanches with other mountain pro-cesses, and their significance for landscape evolution, and Quaternary environmental change, hence;
8. questions of the magnitude and frequency of the landslide events, their time-series properties, and problems of dating them.

4.1. The simple or classic rock avalanche

The classic rock avalanche described in textbooks spreads to an extensive thin sheet, rarely more than 2 to 10 m thick and lobate in plan (Fig. 2). The deposit varies little in thickness and has minor surface relief, commonly with a slight ridge at the distal rim.

This landform is the focus of most modeling efforts (Legros, 2002; Kilburn, 2004), but it describes few actual examples. The classic form arises where runout is relatively unimpeded by local

topography, a condition that applies in only a few cases. With some exceptions, the rock avalanches discussed here also deposited relatively thin lobes of crushed bedrock, but these lobes are only part, commonly a minor part, of the mass. The giant Flims event in the Swiss Alps is an example (Abele, 1974; von Poschinger, 2002). In some instances the classic thin lobate deposit is absent. None has been found in the much-studied Köfels event in Ötztal, Austria, the Hope slide in British Columbia (Fig. 5) and more than a dozen cases in the Karakoram Himalaya.

An important objective of this review is to point out and explore the fact that "rockslide-rock avalanche" is short hand for a diverse suite of mass movements and their deposits. Nevertheless, we stress that all of the events of interest exhibit the characteristics used above to define "rock avalanches." The details of the movement and the final deposit form are diverse, but all of the deposits consist of crushed and pulverized bedrock derived from an elevated source. Huge dense dust clouds commonly obscure what is happening on the ground, but some descriptions suggest an avalanche of rock fragments surging forward, much like a snow avalanche. Sometimes the boulder-covered front and a series of lobes end up far ahead of the main mass. Afterwards, all one sees is a vast field of angular boulders, which gives little sense of the composition and form of the main body of the deposit.

4.2. Complex rock avalanches and intermediate forms

Mass movements commonly are defined in terms of the mechanism of failure (e.g., rotational sliding, toppling) or the rheology of constituent materials. These factors are also important for understanding rock avalanches, but they do not control emplacement forms. Because rock avalanches have great kinetic energy but lack cohesion and tensile strength, topography can concentrate, disperse, or split the debris stream horizontally and vertically, leading to a range of plan forms and surface morphology (Heim, 1932; Abele, 1974; Fort, 1996; von Poschinger, 2002). Topographic blocking by an opposing valley wall or a ridge oriented perpendicular to the direction of movement, can limit or prevent the development of a rock avalanche as, for example, at Köfels and Vaiont in the Alps, and Tsergo Ri in Nepal (Heim, 1932; Heuberger, 1975; Heuberger et al., 1984). In these examples, the landslides stalled before the debris had become sufficiently crushed to stream. Topographic constraints can produce deposits that are tens, evens hundreds, of meters thick (Hewitt, 2002b; von Poschinger, 2002), burying and drastically altering local topography or drainage patterns (Oberlander, 1965; Mathews and McTaggart, 1978; Clague and Evans, 2003; Korup, 2004; Korup et al., 2004, 2006). Large rock avalanches may spread far up and down a main valley. Emplacement at valley junctions or in several adjoining valleys introduces further complexities (Fig. 6).

Of special interest are forms that arise where rock avalanches travel directly across mountain valleys. The thicker part of the deposit may come to rest against the opposite slope (Heim, 1932), or debris lobes may climb hundreds of meters up this slope or even leave the valley altogether (Figs. 7 and 8; Evans, 1989; Evans et al., 1994; Hewitt, 2002a). In some cases, following a steep climb, debris collapses back towards the valley floor, causing a reverse debris stream that may impact the source slope, as in the case of the 1987 Val Pola event in the Italian Alps (Govi

Fig. 5. Photograph of the Hope slide taken soon after it occurred in January 1965. The landslide entrained water and wet, unfrozen sediments on the valley floor, which generated a debris flow that surged down the valley, away from the viewer. (Government of British Columbia photo BC(0)447).
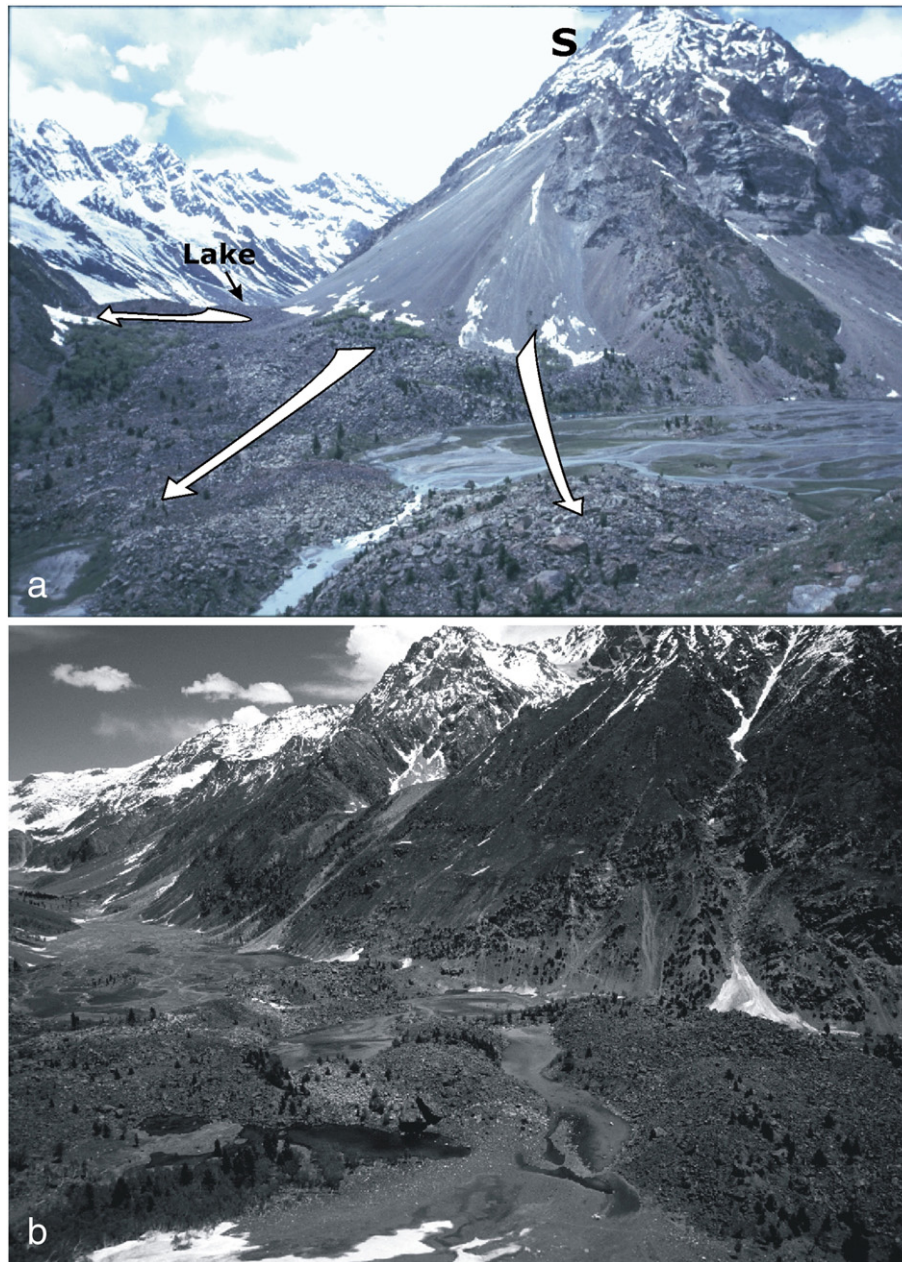
Fig. 6. Debris emplacement at valley junctions by the Naltar Lakes rock avalanche in the Gilgit District, Karakoram Himalaya. (a) Debris derived from the slope at the upper right impounds a lake in a tributary valley. The main debris mass traveled towards the viewing point and split into two lobes (arrowed). The smaller of the two lobes moved to the right and blocked upper Naltar valley; the larger lobe surged to the left another 8 km down the valley. (b) Rock-avalanche deposit viewed from the talus cone below the source slope (S in a). The debris lobe blocks upper Naltar valley. Ridge up to several tens of metres high are both transverse and longitudinal to the flow direction.

et al., 2002; Crosta, 2004). Upon encountering a spur or a slope aligned at right angles to the travel direction, the mobile debris may split into separate lobes that move up and down a valley — the "deformed T-shape" of Nicoletti and Sorriso-Valvo (1991, p. 1370; see also Corominas, 1996).

Rock avalanches can create cross-valley barriers hundreds of meters high that impound large reservoirs (International Strategy for Disaster Reduction, 2000; Korup, 2002, p. 211; Wassmer et al., 2004). They may spread far up and down valleys, such that the resultant dams have broad bases and are resistant to failure (Fig. 9; Costa and Schuster, 1988, p. 10).

Rock avalanches that climb steep slopes leave a distinctive ridge of debris at their culmination. This debris ridge was first described by Heim (1932, pp. 87–89), who called it the brandung, meaning "surge," "breaking wave," or "swash." Because the English terms carry other meanings, we retain Heim's term, brandung (Hewitt, 2002b). The location, form, and height of the brandung depend on the geometric relations of the slope and the trajectory of the debris stream. The highest and most pronounced distal ridges occur where the debris impacts the slope at right angles. Their elevations decrease, and the ridges become less pronounced, as the angle between the slope
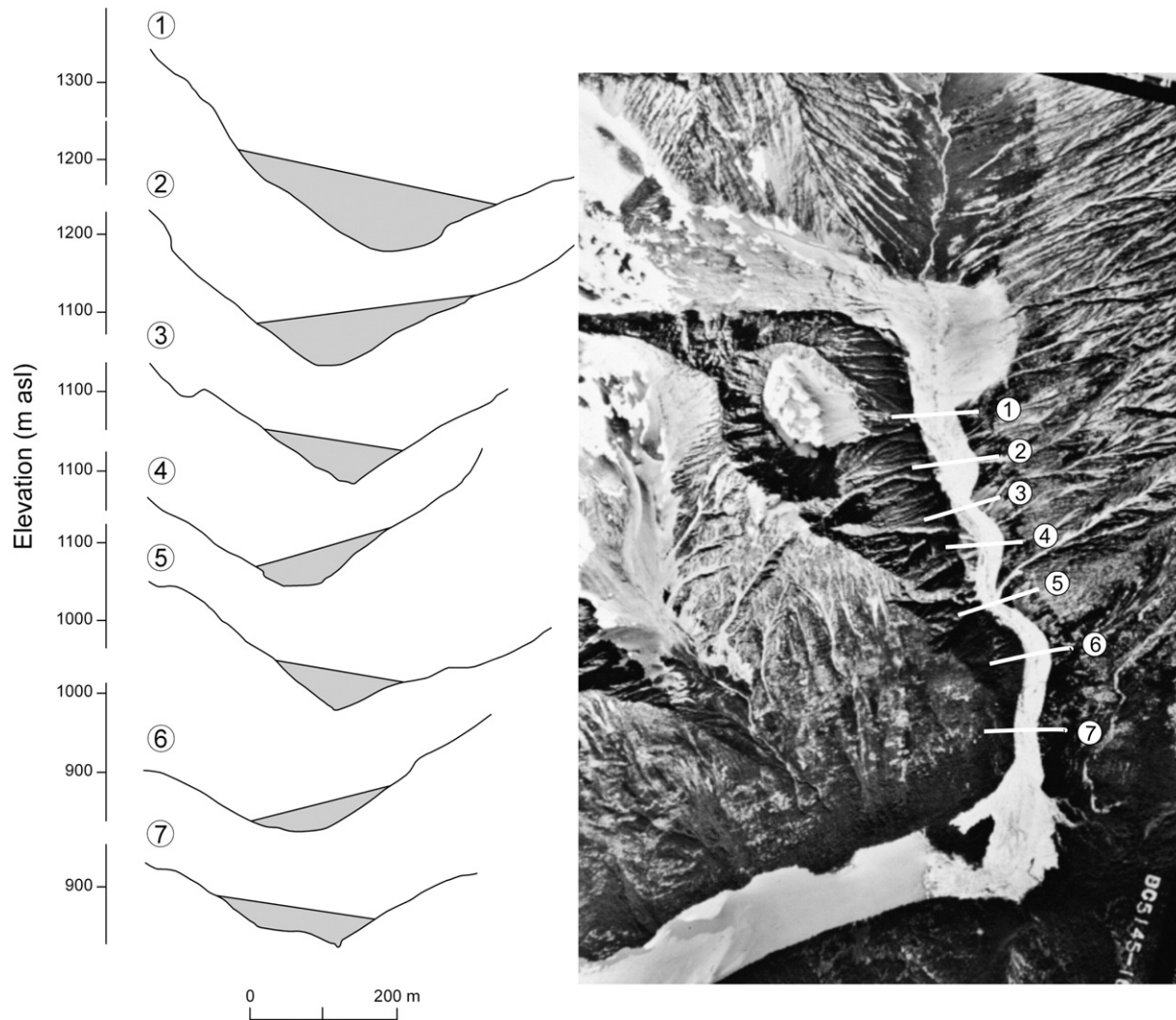
Fig. 7. The Pandemonium Creek rock avalanche, British Columbia, Canada. The landslide had a volume of only 5×10^6 m^3, but travelled over 9 km. The speed of the debris ranged from 80 to 100 m/s as it ran down Pandemonium Creek (Evans et al., 1989).

and the direction of movement decreases. The brandung grades into "caroming" or swash features as movement becomes more closely parallel to the slope (see below). A brandung is superficially similar to lateral moraine remnants, and the two have been confused (Hewitt, 1999).

A rock avalanche that travels down a valley is deflected by valley-side spurs, causing caroming flow (Fahnestock, 1978; Porter and Orombelli, 1980; Evans et al., 1994; Hauser, 2002). In canyons that are narrow compared to the mass of the rock avalanche, a wave-like sequence arises, in which the thickest parts of the flow produce trimlines along canyon walls tens to hundreds of meters above later-arriving materials that stall in the canyon (Hewitt, 2001).

In more open terrain, topographic irregularities may generate substantial transverse or longitudinal ridges in the debris stream and induce interactions among debris moving at different speeds or along different trajectories (Mollard, 1977; Abdrakh-matov and Strom, 2006). Local valley patterns introduce further complications — rock avalanches may exit the valley in which they originate and enter a main valley, cross the mouths of

tributaries, or travel into them (Fig. 10). Some rock avalanches climb over interfluves to deposit material in adjacent valleys (Hewitt, 2002b, 2006b).

In rugged terrain, streaming debris will encounter a variety of topographies. Some parts of the moving debris mass will reach opposing slopes sooner than other parts, setting up complicated stress fields and causing parts of the mass to thicken, thin, collapse, converge, or split into separate streams. Attempts to assess the complex behavior of the Flims landslide illustrate the weakness of two-dimensional models in trying to capture such behavior (von Poschinger et al., 2006). Major differences in rock crushing in different parts of the Flims deposit may relate to convergence and divergence in flow caused by topography. Different crushed units are separated by well defined shear planes (Fig. 11) and are not simply arranged sequentially along the flow path or in simple vertical sequences. Some features are not continuous, suggesting that the moving mass was continually accommodating internal stress. For example, complex differentiation of rock-avalanche facies can be observed in Karakoram examples in sections thicker than about 50 m (see Fig. 4).
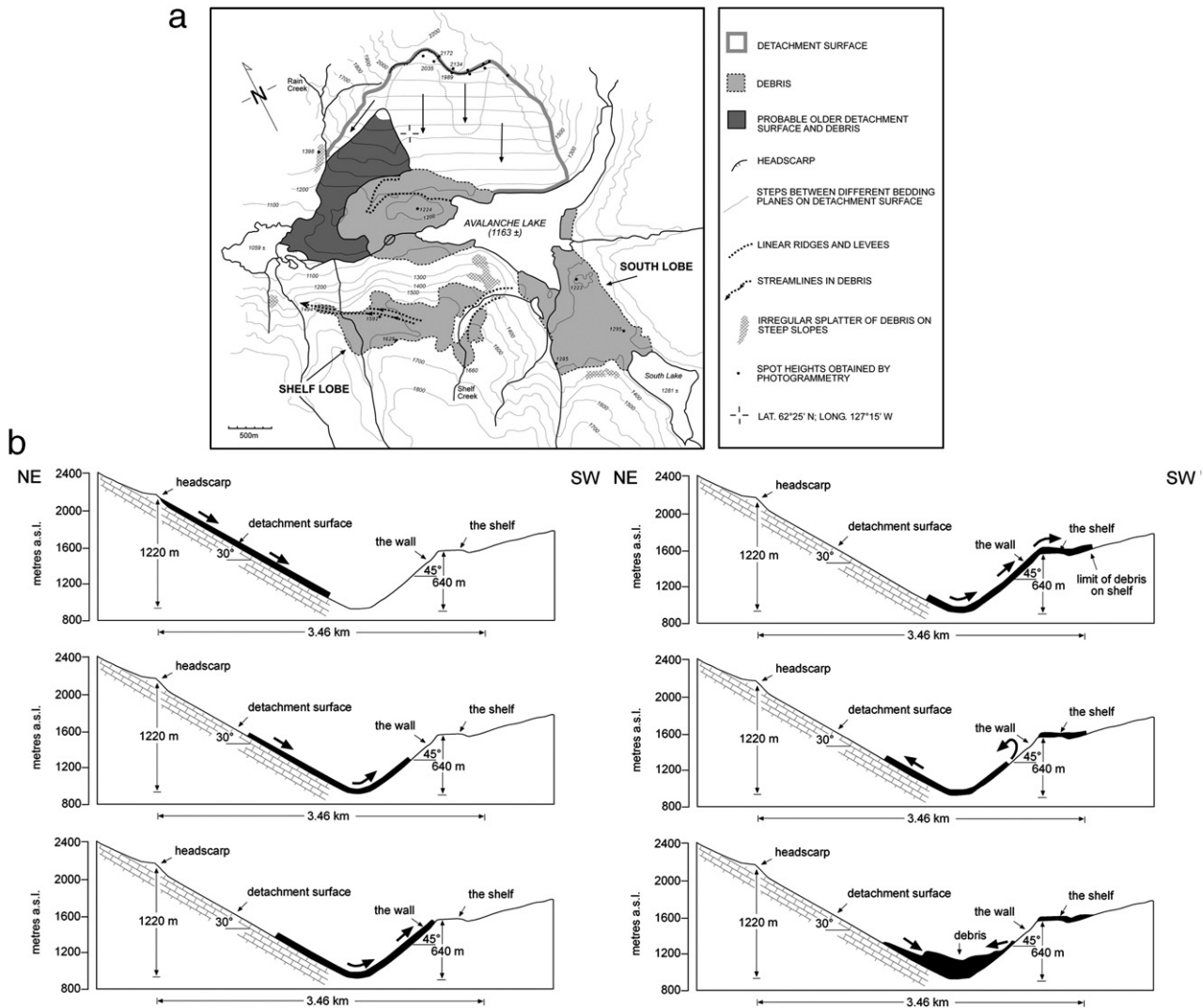
Fig. 8. (a) Map of the Avalanche Lake rock avalanche, Northwest Territories, Canada. (b) Transverse profile of rock-avalanche path, showing sequence of events from failure to emplacement of the debris (modified from Evans et al., 1994).

Such complex events further illustrate the limitations of two-dimensional modeling of rock avalanches, even if the models take into account an opposing slope. Some, if not all of the forms that depart from the classic, thin sheet are possibly intermediate stages of break-up or cataclasis that all rock avalanches go through, preserved as incomplete developments in units stalled by adverse terrain.

In light of these considerations, two end members of sudden, large rock slope failures may be defined. One end member is the classic, thin debris lobe deposited by streaming flow. It is dependent on a high degree of fragmentation during descent and little interference by terrain in the runout zone. The debris spreads relatively unimpeded, affected mainly by its initial mass and the height of fall. The other end member is the translational rockslide. In this case, a large rock mass undergoes limited deformation or disintegration before lodging on the valley floor. Some examples, such as Tsergo Ri, may be very large indeed (Heuberger et al., 1984). The Vaiont slide in Italy is a limiting form of the translational slide; it underwent considerable, but not complete break up before stalling in the valley, and a

rock avalanche did not develop. The 2.2 km³ Waikaremoana landslide in New Zealand has both rockslide and rock-avalanche components (Davies et al., 2006). The rockslide was stopped by the opposing valley wall, but rock avalanches traveled at right angles to the rockslide, up and down the valley.

At several Karakoram sites, the detachment zone of a rock avalanche has also been the source of large translational rockslides. These slides are discussed below as *landslide complexes*.

Recognition of the diverse range of emplacement forms and their distinctive depositional features, to which we will turn in a moment, has opened up new areas of enquiry.

However, topography does not provide the full picture of how rock avalanches may behave during runout.

4.3. Interactions with substrate materials

Complex emplacement forms also arise when rock avalanches descend onto and travel across valleys or basins with deformable sedimentary fills. Longitudinal and transverse

Fig. 9. Deposit of the Lichar-Nanga Parbat rockslide on the Upper Indus River; view up river. The 3-km-wide dam is breached. The rockslide descended from the right and ran up the opposing (west) valley wall (Shroder, 1998).

ridges may develop in the rock-avalanche debris as it travels across these fills. Where the debris moves over deformable substrates, it may transfer stress to them, generating complex folds and faults (Fig. 12). Deformation structures beneath ancient rock avalanches have received some attention and provide important diagnostic features (Krieger, 1977; Yarnold, 1993; Abbot et al., 2002). Holocene examples have been recently described in Norway (Blikra et al., 2006), the Tien Shan (Abdrakhmatov and Strom, 2006), and the Karakoram (Hewitt, 2002b, 2006b).

The responses to substrate conditions are diverse. The Ghoro Choh I rock avalanche in the Karakoram Himalaya, upon impacting the valley, split into a series of longitudinal streams that emplaced five ridges, 10 to 30 m high and up to 3 km long (Fig. 13). This digitate form is relatively rare, although individual longitudinal ridges are common in cases where rock avalanches

Fig. 10. Deposit of the Baltit-Sumaiyar rock avalanche, Hunza Karakoram. The source of the landslide is Ultar Peak (7387 m asl) in the centre background. The rock avalanche descended a glacier and split into several lobes, one of which entered the Hispar valley in the foreground (2200 m asl). The deposit is at least 150 m thick here and longitudinal ridges on the surface are 5–15 m high (Hewitt, 2001).

Fig. 11. Debris of the Flims rockslide — rock avalanche exposed in a quarry in the Upper Rhine Valley, Switzerland. The debris consists of finely to coarsely crushed limestone, but in distinct units separated by shear zones (top right corner). This site is 9 km from the head of the detachment zone near the limit of the debris. The rock avalanche traveled towards the camera.

have crossed valley fills (Clague, 1981); many examples have associated transverse ridges. Prominent constructional landforms develop where large masses of sediment are picked up, transported, compressed, or bulldozed by the landslide (Hewitt, 2004). Sediment units may be dislodged and incorporated into the streaming debris mass. A rock avalanche traveling over saturated sediment may spread farther than it would otherwise, as hydraulic pressure is transferred to the fluid below (Abele, 1997). High-

Fig. 12. Deformed alluvial deposits beneath under the Yarbah Tso rock avalanche, Shigar valley, Karakoram Himalaya. (a) Near-surface folded alluvium exposed by erosion. The block to left of the person was emplaced by the rock avalanche. (b) Contorted and faulted alluvium beneath 10 m of rock-avalanche debris near the base of the source slope. (c) Deformed alluvium in the distal region of the rock-avalanche deposit.
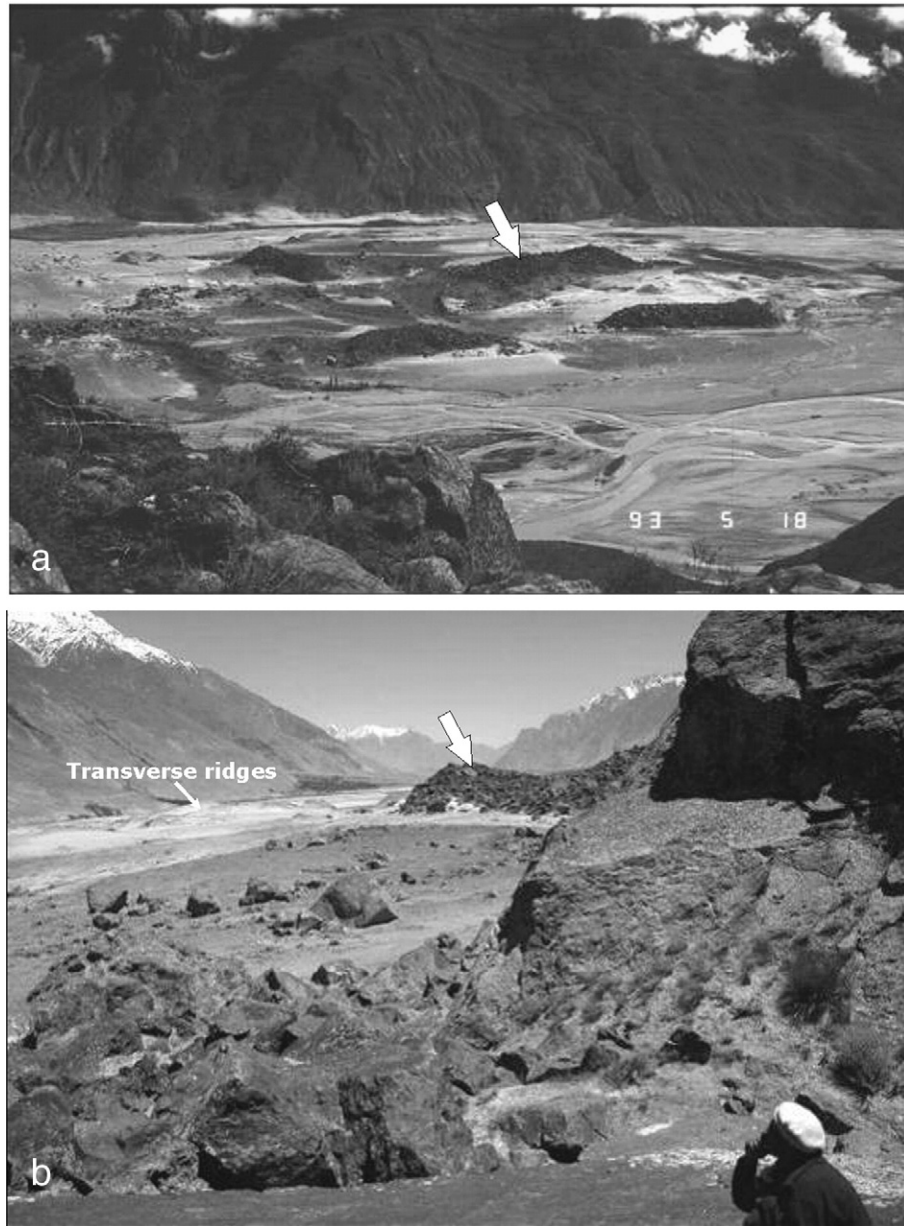
Fig. 13. Longitudinal ridges in the runout zone of the Ghoro Choh I rock avalanche, Shigar valley, Karakoram Himalaya. After plunging into the valley below the source slope, the rock avalanche split into separate lobes that left five ridges 10–30 m high. The ridges consist entirely of tonalite from the source slope, but massive amounts of river gravel and organic and lake sediments were transported between the ridges and bulldozed into sets of transverse radial ridges just beyond them. (a) View of the longitudinal ridges from the detachment zone (one of the ridges is arrowed in a and b). The far side of the valley is 7 km away. (b) Outer margin of the largest ridge, which is 4 km long. Transverse ridges of river gravel can be seen in the distance beyond, the longitudinal ridge.

velocity, fluidized lobes may extend up and down-valley, far beyond the main deposit (Fig. 14; Orwin et al., 2004).

Streaming rock-avalanche debris may also split vertically into two or more sheets. While part of the mass slices into and beneath softer valley fill, the remainder passes over the fill. In other cases, masses of rock-avalanche debris separate from the main lobe and travel over a water-saturated substrate some kilometers beyond the limit of the main, continuous deposit, evidently at a great speed (Fig. 15). This phenomenon has been documented for the Flims and Fernpass events in the European Alps, where the term 'toma' has been applied to the resulting landforms (Abele, 1997; von Poschinger, 2005), and

the Naltar Lakes rock avalanche in the Karakoram (Hewitt, 2002a, b). In these examples, conical mounds and small steep hills of pure rock-avalanche material occur in isolation kilometers down-valley from the continuous rock-avalanche deposits.

It is remarkable that rock-avalanche material rarely mixes with entrained and tectonized substrate material, except in a thin, complex, mixed zone at the base of the flow. In those cases where substantial mixing does occur, the character and mechanism of the mass movement itself changes through a large uptake of moisture or entrainment of wet sediment (Pflaker and Ericksen, 1978; Hauser, 2002).
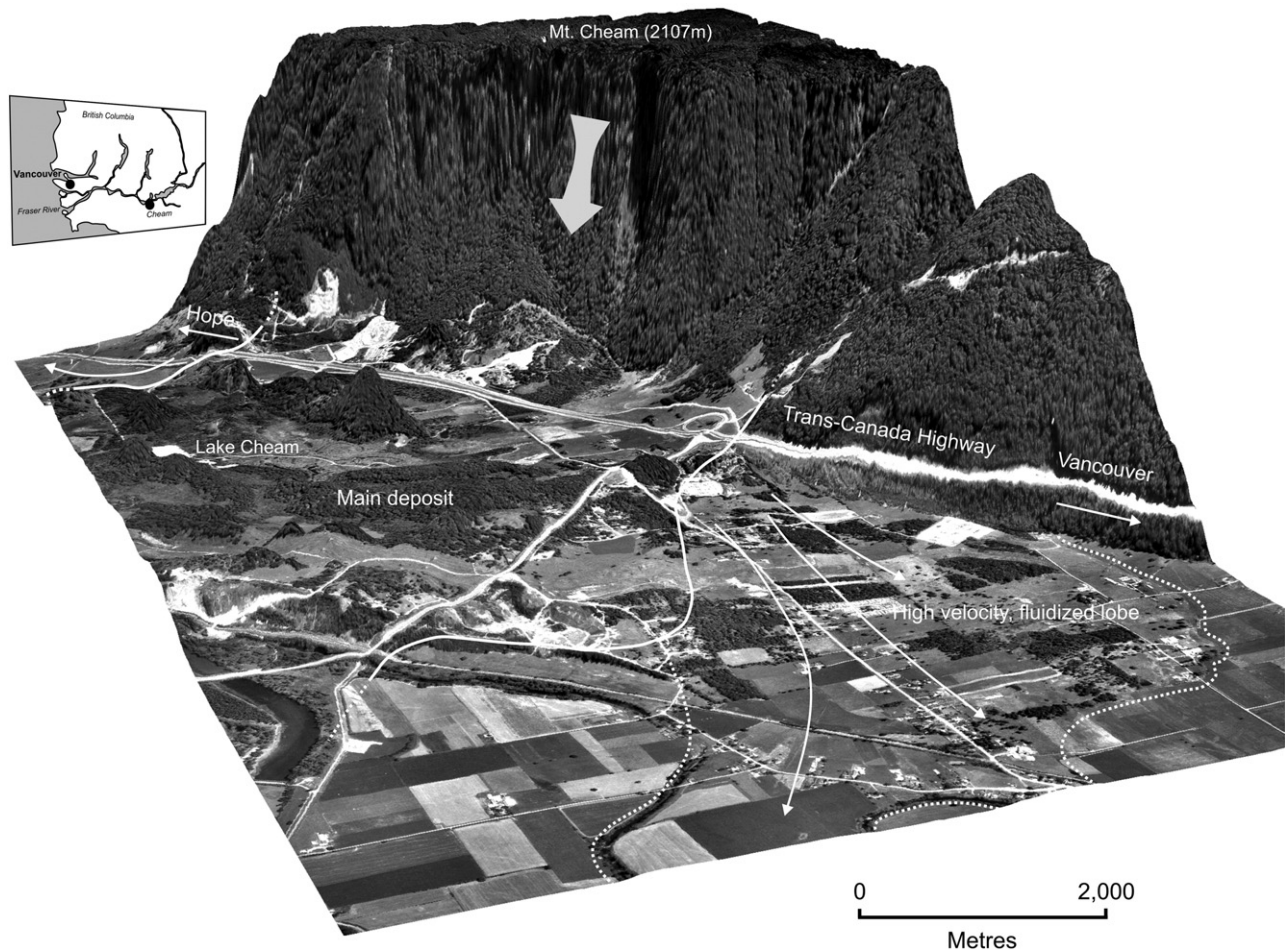
Fig. 14. Liquefaction and fluidization explain the marked difference in morphology between the eastern and western parts of the Cheam rock avalanche, British Columbia, Canada (Orwin et al., 2004). The east half of the deposit comprises steep-sided, arcuate ridges with intervening flat-bottomed depressions. Some of the ridges rise several tens of metres above the surrounding terrain. The west half of the deposit is lower and has a subdued, gently rolling surface, which slopes, on average, 1.5° to the west (maximum = 19°). This part of the deposit extends from the higher elevation, eastern region across a low terrace and onto the Fraser River floodplain, where it terminates in an abrupt arcuate front. (Reproduced with permission of Springer Science and Business Media.)

4.4. Interactions with glaciers

An area of landslide research that is attracting increasing attention is the relation between mountain glaciers and catastrophic rock slope failure. Two separate issues are of interest:

1. the impact on slope stability of the loss of snow and ice in rapidly deglacierizing mountains, and
2. the effect of rock avalanches on glaciers and glacier behavior, and the fate of rock-avalanche debris emplaced on glaciers.

The twentieth century was marked by global climate warming (Houghton et al., 1996) and by loss of snow and glacier ice in most mountains. The mountains of western North America have lost about one-third of their ice since the late nineteenth century. The loss is most conspicuous in the high mountains of southeast Alaska, southwest Yukon, and British Columbia. For example, more than 1000 km^3 of ice has disappeared from Glacier Bay, Alaska, since George Vancouver's visit in 1793, and at least half of the loss has occurred in the past 150 years (Clague and Evans,

1994b). A comparable reduction in glacier cover has been documented in the Karakoram and most other Himalayan and Inner Asian mountain systems (Shroder, 1993; Calkin, 1995).

Some researchers have suggested that twentieth-century ice loss has debuttressed steep unstable alpine slopes and contributed to their catastrophic failure (Evans and Clague, 1988; Bovis and Stewart, 1998; Deline, 2005). Certainly, many historic rock avalanches have sources on slopes that were supported by glacier ice until recently (Geertsema et al., 2006). The possible association of glacial debuttressing and catastrophic rock slope failure raises questions about the frequency of large failures in high mountains later in this century as glaciers continue to retreat.

In glacierized mountains rock avalanches fall, and leave their deposits, on glaciers (Figs. 16 and 17; McSaveney, 1978, 2002; Evans and Clague, 1988; Jibson et al., 2006). A variety of important issues arise from the behavior of rock avalanches that travel over glaciers, the deposits they emplace on glaciers, and subsequent glacier responses (D'Agata et al., 2004; Hewitt, 2005). The blankets of debris, although thin, have an important effect on glacier regimen by insulating ice in the ablation zone. They may even trigger surges (Tarr and Martin, 1912; Gardner

Fig. 15. A small mound of rock-avalanche debris, or 'toma,' associated with the Fernpass rock avalanche in the Austrian Alps. The mound is one of several that are at least 3 km beyond the terminus of the main deposit.

and Hewitt, 1990; Hewitt, 2006b; in press). Ultimately, glaciers move the debris blankets to their margins where they may be deposited as moraines (Kirkbride, 1995). Interpretation of these moraines as the products of past climate conditions may, therefore, be problematic (Larsen et al., 2005). Studies of the Brenva Glacier, Italy, have shown that several rock avalanches traveled over and were deposited on the glacier in the past century. Fluctuations of the glacier were found to be out-of-phase with climate changes and, instead, reflected repeated burial and reduced ablation beneath the landslide debris (Valbusa, 1921;

D'Agata et al., 2004; Deline, 2005). Much of the finer fraction of the rock avalanche debris may be entrained by meltwater streams and transported down valley. Dust may be removed by wind and transported to distant regions. It seems likely that much of the debris on glaciers in Alaska and the Karakoram has been emplaced by landslides, and glacial transport of this vast amount of debris is a major contributor to mountain denudation and sediment transfers.

From a study of 17 events in western Canada, Evans and Clague (1988) concluded that travel over glaciers generally
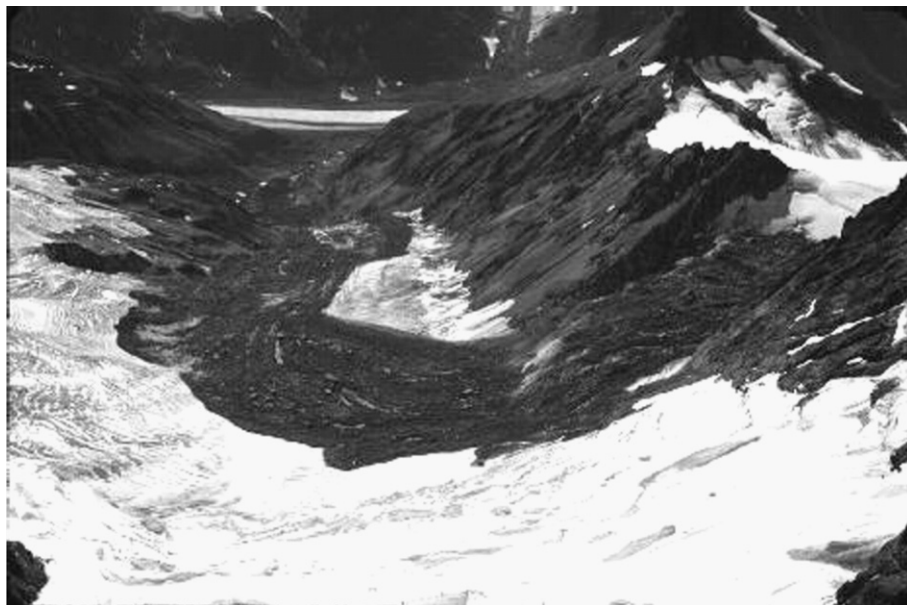

Fig. 16. The 1990 Frobisher Glacier rock avalanche in the St. Elias Mountains in northwest British Columbia. This rock avalanche fell onto and flowed down Frobisher Glacier. Note the secondary lobes of highly fluid debris at the far left that ran away from the main debris body.

Fig. 17. Sheet of rock-avalanche debris, about 1.5 m thick, covering Black Rapids Glacier (S is source). The rock avalanche was triggered by the Denali earthquake (magnitude 7.9) in November 2002.

increases rock-avalanche mobility. Singular catastrophes can occur where these landslides travel on and beyond glaciers to inhabited areas (Pflaker and Ericksen, 1978; Hauser, 2002).

4.5. Compound events and landslide complexes

Entrainment of substrate material or moisture may transform a rock avalanche into a different type of mass movement. Rock avalanches that cross or travel down flood plains or glaciers can mobilize so much moisture and wet sediment that they become debris flows (Fahnestock, 1978; Pflaker and Ericksen, 1978; Abele, 1974, 1997; Fauque and Tchilinguirian, 2002; Evans et al., 2001). Massive rock slope failures in humid mountains or during wet seasons tend to readily convert into debris flows. Entrained plant debris from forested slopes further complicates flow behavior (Schuster et al., 2002).

The Ghoro Choh I event in the Karakoram, mentioned above, illustrates all these complications. Beyond the enormous longitudinal ridges of rock-avalanche debris is an apron of crushed rock mixed with flood-plain alluvium. The apron covers an area of more than 5 km^2 and thins from about 3 m just beyond the rock-avalanche ridges to a few centimeters at its distal edge (Hewitt, 1999). It was deposited by a debris flow that formed when the rock avalanche encountered very wet alluvium or a river. However, large blocks of the distinctive rock-avalanche lithology traveled with or over the debris flow to the distal rim. In places, the blocks lie beyond any discernable debris-flow deposits. Abundant rockslide and rock-fall debris litter the area surrounding the detachment zone, demonstrating that the Ghoro Choh event, formerly classed as a moraine, is a postglacial landslide (Hewitt, 1999). A radiocarbon age on a piece of detrital wood in the landslide debris indicates that the event is no more than 8000 years old.

Rock-slope detachment zones, like runout forms, exhibit considerable variety. Rock avalanches are rarely the only process involved in large-scale failure of steep rock slopes. Large bodies of rock may fail and move downslope, but not break up sufficiently to generate a rock avalanche. Examples in the Karakoram include the Yashbandan-Barut, Gomboro, and Upper Henzul failures. In addition, a slope may generate countless rock falls before it fails catastrophically, and rock falls may continue for years, even decades, after a large rock avalanche. The latter has been observed at Bualtar Glacier in the Karakoram following events in 1986, and to the present time from the detachment zone of a late 19th century rock avalanche (Hewitt, 2006b). Similarly, McSaveney (2002) showed that the slopes that produced the two Mount Fletcher rock avalanches in New Zealand generated rockfalls for at least 50 years before they fell.

Of special interest are sites where more than one catastrophic failure has occurred and sites affected by slow, deep-seated, rock-mass creep (Fig. 18). Features of interest include incipient or minor failures and sagging (sackung) that define large, detached or partly detached rock masses discussed below. Some of these rock masses may fail during large earthquakes or extreme weather events and develop into rock avalanches. On the other hand, there are numerous examples of centuries-old, unstable rock slopes that have not failed catastrophically. Detailed studies of some of these slopes have shown that they present no unusual risk (Thompson et al., 1997; Merrien-Soukatchoff and Gunzberger, 2006). As a result, it remains uncertain whether sackung is a necessary precondition for catastrophic collapses.

This discussion leads us to propose two distinctive contexts for large landslide complexes. On one hand, there are sites with single events involving two or more different types of mass movement. On the other hand, there are sites where two or more
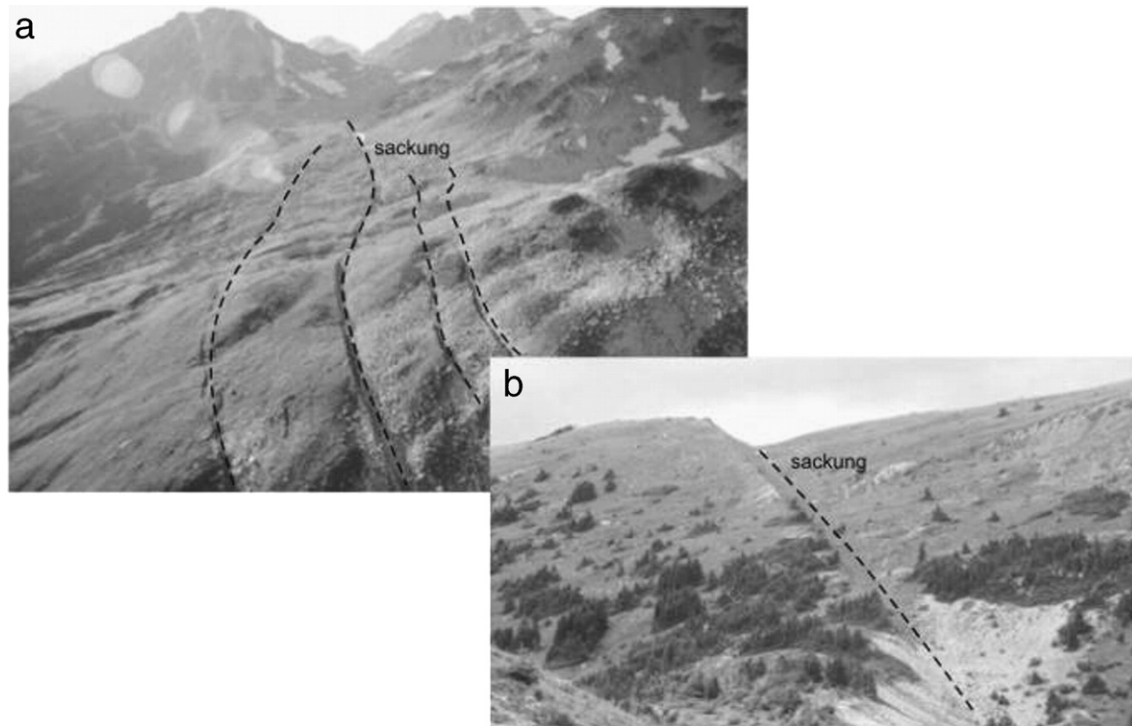
Fig. 18. Antislope scarps and trenches produced by slow creep of large rock masses (sackung) in the Coast Mountains of British Columbia. (a) Lillooet River valley, north of Pemberton. (b) The Hell Creek "fault" in the Bridge River valley. Movement in these two cases has occurred preferentially along the gouge zones of Tertiary faults.

distinct landslide masses, possibly involving different mass movement processes, originate from the same detachment zone. More than one type of mass movement is found in both of these contexts; thus existing descriptive and mechanical classifications have limited value in describing and understanding the
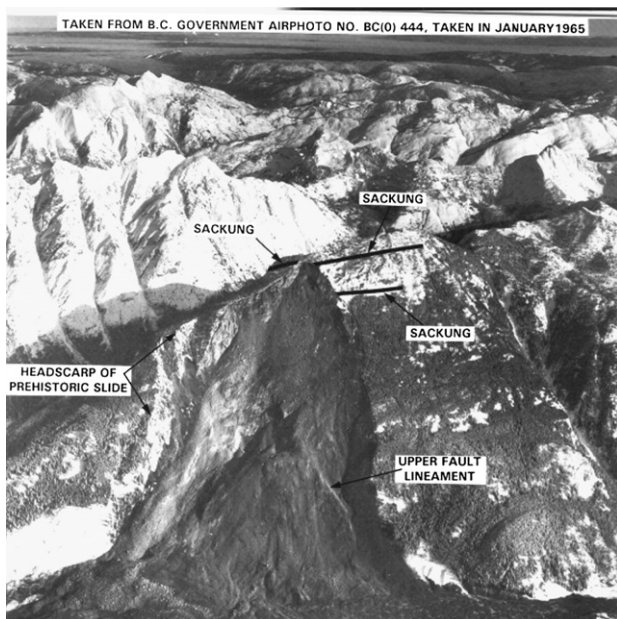

Fig. 19. Source zone of 1965 Hope slide showing remnants of an early Holocene landslide scarp and, to the right, sackung lineaments. (Government of British Columbia photo BC(0)444).

landslides. The important point for mountain geomorphology and landslide hazards is to emphasize the presence of two or more processes and forms, and how that complicates the erosional and depositional record.

It seems appropriate to describe both these cases as "landslide complexes" (Hewitt, 2004). It is difficult to assess the extent of such complexities in high mountains because of difficulties of access and erosion of the more accessible deposits. It thus may be some time before we can determine the full significance of large landslides in some mountain regions.

4.6. Relation between sackung and catastrophic slope failure

Many rock avalanches occur on slopes that have slowly deformed for thousands of years or more. Well known examples include the Hope and Frank slides in western Canada (Fig. 19). Sackung occur above the detachment zones of many rock avalanches in the Norwegian fiordlands (Blikra et al., 2006), and have been found above the Upper Henzul, Nomal, and Gol-Ghone landslides in the Karakoram Himalaya (Hewitt, 2006b). Slow, deep-seated deformation at these sites is indicated by surface features typical of sackung (Zischinsky, 1968), including bulging of the toe of the deforming rock mass; multiple, more-or-less, slope-parallel scarps on the upper part of the slope; and ridge-crest grabens. Detailed examination of the geomorphology of the slopes and trenching of sediment fills in depressions behind antislope scarps typically show a complex movement history, in some cases with phases of movement separated by intervals with little or no activity (McAlpin, 1995; Thompson et al., 1997).

An important question is whether lengthy, slow, deep-seated movement of a slope necessarily leads to catastrophic failure. The answer seems to be no. Sackung with movements spanning the past 10,000 years or more are extremely common in many mountain ranges (e.g., the Coast Mountains of British Columbia; Bovis and Evans, 1996), yet few are associated with rock avalanches. Further, geotechnical analysis of many sagging slopes suggests that they are unlikely to fail catastrophically, even if shaken by an earthquake. On the other hand, slope sagging does indicate a dilated, closely jointed rock mass, which may be a secondary contributing factor to some large rock avalanches and a major factor in a few regions because of relations between lithology and erosional history (Blikra et al., 2006).

Slopes on which rock avalanches have occurred, or will occur, require a special set of topographic, lithologic, and structural conditions that predispose them to rapid, deep-seated failure. For example, the 1965 Hope slide (47×10^6 m^3) in southwestern British Columbia occurred on a slope underlain by closely jointed greenstone intruded by dykes or sills of altered felsite dipping parallel to the slope. Part of the basal rupture surface was one of the felsite sheets. Had the felsite been absent, the landslide might not have occurred, in spite of the fact that the slope had been slowly sagging throughout postglacial time (Savigny and Clague, 1992). The famous 1963 Vaiont landslide (ca. 200×10^6 m^3) also happened on a slope with a long history of sagging. In this case, the catastrophic failure resulted from the filling and lowering of a reservoir. It is uncertain that catastrophic failure could have occurred if the Vaiont dam had not been built (Ghirotti, 2006).

5. Rock avalanches in the geologic record

Study of rock-avalanche deposits in the geologic record has been beset with problems, but considerable progress has been made in their identification and interpretation in recent years. Major contributions have come from research in tectonically active areas of the United States, notably California and Arizona (Krieger, 1977; Yarnold and Lombard, 1989; Topping, 1993; Friedmann, 1997; Abbot et al., 2002). Developments have benefited from key studies of comparable, late Quaternary rock avalanches in the same region, for example the Blackhawk and Martinez Mountain rock avalanches in California (Shreve, 1968; Bock, 1977; Johnson, 1978). For both these and more ancient events, investigations have turned on the ability to recognize diagnostic sedimentary features that distinguish rock-avalanche deposits from other coarse, unsorted or poorly sorted materials in the same environments.

Few of the ancient events described in the literature seem to approximate what we have termed the classic rock-avalanche form. Some of the first discussions of complexities related to topography and terrain, and to runout on sediment fans or valley fill and into lakes, came from stratigraphic and paleogeographic investigations (Krieger, 1977). Interactions with wet, mobile, and erodible sediments, including features described by the phrase "landslide-tectonised" substrates, as well as transformation of rock avalanches into debris flows, were first recognized through investigations of the rock record (Topping, 1993;

Yarnold, 1993). Stratigraphers also had to deal with the problem of defining the origin and margins of landslide deposits that had been extensively eroded before burial or after exhumation (Friedmann, 1997; Abbot et al., 2002). It can be difficult enough with late Quaternary examples to connect distal remnants of landslide debris with their source slope; in the case of ancient landslides, the original mountains or mountain fronts may have disappeared millions of years ago (Topping, 1993). A particular value of the stratigraphic and sedimentological research on the rock record has been to draw attention to features that are not readily apparent in historic examples and relatively undisturbed deposits.

For stratigraphers, problems in reconstructing ancient events are almost the opposite of those facing geomorphologists working in the present and recent past. In the case of the latter, interpretive efforts have been disproportionately influenced by the vast, boulder-covered surface and excessive runout of rock avalanches. In the ancient record, cross-sections and parts of the main body of the deposits are more likely to be encountered. Outcrops of basal and lateral margins and other features provide exposures that are rarely available to those working on recent deposits. As a result, stratigraphers have paid greater attention to the internal sedimentary characteristics of rock avalanche deposits, which are important in themselves and draw attention to aspects of landslide dynamics that had been absent from, or their significance missed, in present-day process studies and rock-avalanche models.

Initially, research focused on identifying a distinctive rock-avalanche material (Hewitt, 1999). Subsequently, stratigraphers recognized that some of the major sedimentary problems were not posed by the rock avalanche itself. Much as is argued here, a whole range of interactions with opposing terrain and variable substrates were considered, with climate, regolith, and other surface processes playing significant roles. These interactions not only complicate the landslide process but mask the nature and central role of the rock avalanche. In addressing these problems, stratigraphers came to recognize a distinctive three-dimensional architecture of sedimentary bodies associated with catastrophic slope failures (Yarnold and Lombard, 1989).

Rock-avalanche deposits can be defined and differentiated in terms of three vertical elements and two horizontal elements. The vertical elements comprise: the (1) subaerial surface and upper zone or carapace, which is dominated by coarse blocks; (2) the main body of the deposit, in which coarser fragments are embedded in crushed and pulverized matrix and the abundant "powder" of Heim (1932); and (3) the singular "basal zone" of Yarnold and Lombard (1989), where interactions with substrate materials are important (Johnson, 1978; Abbot et al., 2002). The two horizontal elements are the main body of the rock avalanche deposit and the distal rim.

Deposit characteristics identified by stratigraphers apply to most, if not all, rock avalanches or, more exactly, have come to provide a set of criteria definitive of their place in the sedimentary record. However, as we have seen, the failed rock mass is transformed as it descends from the detachment zone, runs out, and comes to rest. The transformations give rise to a range of distinctive features in the sedimentary record, over and above

the distinctive properties of the rock avalanche deposit. Parts of the failed mass may remain largely intact as a translational slide or block glide, or become fractured with minimal relative displacement of blocks to form crackle-, or jigsaw-brecciated units (Shreve, 1968; Laznicka, 1988). The debris pile may include large brecciated units amid thoroughly crushed and distorted material (Hewitt, 2002b, his Fig. 5b, p. 352). The main body may be hundreds of meters thick, yet completely crushed and pulverized, while some lobes disperse to a thin sheet as in the classic rock-avalanche form. Remnants of such interactions have been confirmed in the geologic record (Krieger, 1977; Abbot et al., 2002).

What separates ancient deposits from most of the examples discussed here is their preservation in the geologic record. Why do many, perhaps thousands, of rock avalanches survive from distant geologic eras in the southwestern United States? Preservation relates mainly to a distinctive structural and tectonic history. The likelihood of deposit preservation is greater in extensional and trans-tensional tectonic environments, such as the U.S. Basin and Range and the Central Volcanic Zone on the North Island of New Zealand. Known events in the southwestern United States all contributed to sedimentation in actively subsiding basins between faulted mountain blocks, as recognized more than 50 years ago by Longwell (1951), who described them as "...megabreccia developed downslope from large faults..." His "megabreccia" refers specifically to poorly sorted, lithified sediments with coarse, angular fragments, including large blocks. No doubt, similar situations exist in Inner Asia, for example, where many ancient rock avalanches remain to be discovered in the sedimentary record. In contrast, in compressional tectonic regimes, including the three regions we draw heavily on, there is little likelihood that the landslide deposits will be preserved in the geologic record. Indeed, if our interpretations are at all correct, millions of Quaternary catastrophic landslides have been completely removed by erosion in these areas. Where rock avalanches occur in mountains with rapid uplift and actively incising rivers, the chances of long-term preservation are very low.

5.1. The climate connection

Aridity reduces the chance of complete transformation of a rock avalanche to a wet flow and also aids in recognition by reducing vegetation cover. It also greatly increases the likelihood that rock-avalanche deposits will survive at the surface, irrespective of topographic and structural conditions.

Conversely, the defining characteristics of rock avalanches and their deposits apply in every climate where they occur, from equatorial high mountains to the flanks of nunataks at high latitudes; from coastal mountains to the arid Inner Asian ranges; and even on Mars and the Moon. They apply throughout our three regions, despite the fact that these regions exhibit enormous topographic and climatic diversity. Even at the event scale, some rock avalanches travel from snow- and ice-clad slopes to arid valley floors, or from barren alpine slopes to coastal rain forest. The altitudinal organization of biophysical environments and "cascading erosional systems" of mountains

are integral to understanding what individual rock avalanches do and what happens to them (Owens and Slaymaker, 2004). They do not, however, determine whether the mass movement is a rock avalanche.

Rockslide — rock-avalanche processes and materials are essentially "aclimatic" or "azonal." What is definitive about them is found in all climate zones — like rock fall but unlike climate-limited processes such as solifluction. Certainly, landslides may be triggered by climate events and their incidence differs greatly in different climatic zones. But climate has no definitive influence on the rock-avalanche process or the properties of the deposits.

In the context of climate, the key distinction is between the original landslide mass and what can happen to it in different environments. Direct and indirect climatic influences are critical to: the fate of the landslide once it occurs; its role in the landscape; whether, in fact, it will enter the record as a "rock avalanche" or survive at all. Attendant landforms and other processes operating in the same area depend on climate. Indeed, if these were not so important, much of the discussion of complex runout and emplacement here would be redundant. The chance of a rock avalanche occurring, and the range of attendant complications when it does, vary greatly with climate. Rock-wall failures in humid mountains or wet seasons are much more likely to be converted to wet flows, and entrained biomass from heavily forested slopes can further complicate behavior and modify deposition (Schuster et al., 2002). As described earlier, rock avalanches in glacierized terrain can be modified by the presence of abundant snow and ice, by travel over glacier surfaces, and by the morphological features of glaciated valleys. The legacies of glaciation and deglaciation are key factors in rock-avalanche incidence and behavior in all three regions emphasized here and, ultimately, depend on climate change. Glaciation increases the relief between ridge crests and valley bottoms, and deglaciation debuttresses steep, unstable valley walls. Along with vertical shifts in climate generally, glaciation and deglaciation affect the distribution, magnitude, and frequency of rock avalanches and their roles in the landscape.

6. Landslide time series and dating

In the spectrum of Earth surface processes, rock avalanches are relatively uncommon events. Most of those that are known are prehistoric, but because many rock avalanches have been misinterpreted as products of other processes, their frequency is greater than is normally assumed (Hewitt, 1999; Larsen et al., 2005; Blikra et al., 2006). In addition, many large historic landslides in remote mountains are not observed or go unreported, biasing opinions about rock-avalanche incidence. These issues indicate the need for direct risk assessment, but a reliable assessment is hampered by difficulties in analyzing the stability of high-relief, steep rock walls and in working in otherwise inhospitable terrain. For these reasons, dating prehistoric events assumes special importance. Where there are substantial and fairly representative sets of known events, it may be possible to generate time series that will indicate the kinds of temporal and spatial distributions involved.

In this respect, a number of models may be applied to interpret the magnitude and temporal and spatial frequencies of rock avalanches. Models focused mainly on triggering mechanisms tend to represent catastrophic rock slope failures as rare, essentially random responses to extreme seismic or weather events. An alternative model, relevant to the postglacial period, assumes a temporal relaxation or "exhaustion" process. The number and size of events decline as unstable sites, created by, say, glaciation or rapid incision, are used up (Cruden and Hu, 1993); a model not unlike a "paraglacial response" (Church, 2002).

However, geomorphic environments in high mountain ranges are organized vertically, and environmental change tends to occur by vertical zonal shifts. A consequence of this zonation is that different magnitude-frequency distributions can apply simultaneously in different elevation bands (Hewitt, 1993). Rapid uplift and incision appear to maintain steep rock walls in the Karakoram and adjacent ranges (Burbank et al., 1996), the Southern Alps of New Zealand (Whitehouse, 1988; Prebble, 1995), and the St. Elias Mountains (O'Sullivan and Currie, 1996; Spotila et al., 2004). If the exhaustion model is applicable to mountain ranges in active orogens, it would seem to compound or overprint a more fundamental rate process dependent on the relation of uplift to erosion. In the long run, landslide incidence can be expected to approximate a steady state if uplift and denudation are more or less balanced.

The probability distribution of landslides is altered by changes in any of the processes that control the stability of slopes, notably glaciation, extreme weather, and seismicity. These changes may be difficult to recognize when there are only limited time sequences to work with. Preservation bias towards younger events and assumptions of record completeness are significant issues in estimating rock-avalanche frequency, even in regions with relatively good spatial and temporal dating control. Using a dataset of 42, largely prehistoric events, Whitehouse and Griffiths (1983) estimated the frequency of large ($>10^6$ m^3) rock avalanches in the Southern Alps of New Zealand to be one per century. However, this estimate assumes that the most recent part of the record is complete. McSaveney (2002) found that this assumption is invalid because it does not account for the many rock avalanches that fall onto glaciers and for which no evidence remains. He determined rock-avalanche frequencies using historic rock-avalanche records to one event per 20–30 years, but even this value is now known to be a gross underestimate of the true frequency (M. McSaveney, written communication, 2007). The implication here is that spatial and temporal frequency estimates for rock avalanches are likely to be mis-representative, even with reasonable sample sizes. Resolving these issues depends on obtaining adequate, temporally representative sets of rock avalanches and, in turn, reliably dating them. In some orogens, all moderate and large landslides can be precisely located using networks of modern seismometers. For example, locations of rock avalanches in the Southern Alps now can be pinpointed in real time to less than 1 km.

Application of new or improved techniques is adding to and, in many cases, revising our knowledge of the late Quaternary record. Of the available dating techniques, surface-exposure cosmogenic nuclide dating is most likely to significantly change how we view the spatial and temporal nature of rock avalanches. Obtaining accurate ages on rock avalanches using this technique requires that the dated blocks have no nuclide inheritance from prior exposure and that the deposit has not been exhumed or significantly eroded (Gosse and Phillips, 2001; Colgan et al., 2002). Cosmogenic nuclide dating is particularly suited to dating these events because catastrophic failure exposes large amounts of previously buried rock at the time of failure. Furthermore, many rock-avalanche deposits survive for millennia in the landscape and, in the absence of further failures, are stable. These attributes satisfy the main caveats for obtaining reliable surface-exposure cosmogenic ages (Gosse and Phillips, 2001). Continued refinements in estimates of productions rates of cosmogenic nuclides are further improving the accuracy of surface-exposure dating (Kubik et al., 1998; Kubik and Ivy-Ochs, 2004). Another significant development is the potential use of other cosmogenic nuclides to date landslides. ^{14}C, for example, accumulates in situ in quartz and could be used to accurately date young surfaces if an extraction technique for the isotope could be developed.

Wider application of cosmogenic dating will generate new insights into the relations between the frequency of rock avalanches and orogeny and glaciation. Dating of previously "undateable" rock-avalanche deposits and re-dating of deposits whose age is disputed may lead to significant reinterpretations of landslide frequency-magnitude relations. For example, coarse debris at the foot of Beinn Alligin, Scotland, previously described as a rock-glacier deposit or the deposit of a rock avalanche redistributed by glacier flow at the end of the last glaciation, was re-dated using cosmogenic 10Be to 3950±320 years BP and thus was not associated with a glacier (Ballantyne and Stone, 2004). The power of surface-exposure cosmogenic dating lies in resolving temporal changes in rock- avalanche frequency and the relation between triggers and mountain-range evolution in areas where traditional dating techniques cannot be applied (Hermanns et al., 2001). The ability to use surface exposure dating on a wide range of rock types is important in regions such as the Karakoram and St. Elias Mountains where there are large numbers of rock avalanches, but only a limited number of event ages. Better spatial and temporal control on rock-avalanche frequencies may, therefore, lead to new ideas on the role these events play in landscape development.

7. Geologic and erosional predesign

Many factors affect the development of large rock slope failures, from the lithology of the failed rock slope to the topography of the runout zone. They give rise to a range of possible sequences and forks in the development of the landslides. Most of the factors have been investigated separately in specialized studies; here we review them with a focus on how they relate to the entire landslide event.

Lithology, and tectonic, structural, and erosion history are the fundamental or root causes of rock slope failure. It is useful, as Wieczorek (2002) said, to distinguish these root causes from landslide triggers, which determine the actual moment that an
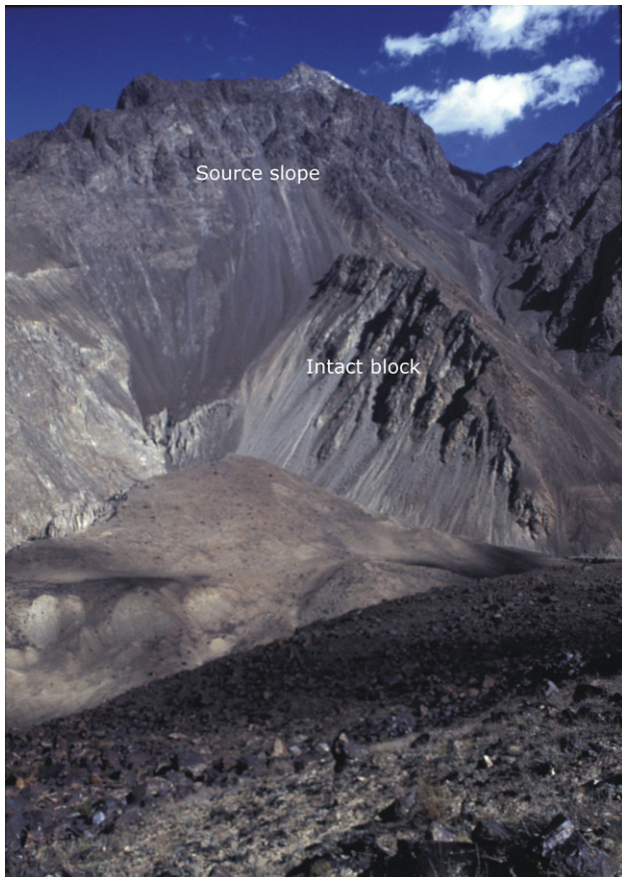
Fig. 20. Satpara Lake-Skardu rockslide — rock avalanche, central Karakoram. A huge, largely intact block of rock descended from the detachment zone of the rock avalanche. The rock avalanche traveled across the valley and left a brandung near the photo point, some 350 m above the intervening valley floor.

event occurs. The most important triggers are earthquakes and extreme weather. It seems that the processes that give rise to steep rock slopes and determine their stability at any given time combine with the nature and magnitude of the triggering events to determine the size and geometry of the failure. Thereafter, topography, substrate, and other conditions in the travel path affect the style of descent and what happens to the disintegrating rock mass.

One view of cause is that landslides occur at sites that are pre-determined by tectonics, lithology, and terrain development to fail. Geotectonic history establishes the potential form and scale of failure, or its "predesign," to use a term from Scheidegger (1998) (see also Weidinger, 2004). Predesign is a general concept encompassing and influencing the erosion of mountain valleys in the broadest sense. It suggests that the morphology of valleys and mountains is preconfigured by tectonic events and stress fields that create weaknesses in crustal materials (faults, jointing, bedding, and foliation) that, in turn, set the stage for rock-slope failure (Fig. 20). Yet, the word "predesign" hides many processes that can lead to a range of distributions over time.

Topography reflects distinctive subaerial conditions and environmental histories that are controlled by tectonic and other independent factors. These factors dictate where and when

slopes fail catastrophically and how a failed mass will behave. As noted in many formerly and presently glacierized mountains, glacial steepening and debuttressing are important mechanisms in preparing slopes for catastrophic failure (Evans and Clague, 1988; Holm et al., 2004). Recently, many catastrophic land-slides have been identified originating on the walls of coastal fiords, their deposits hidden beneath the waters below (Boe et al., 2003; Braathen et al., 2004; Blikra et al., 2006). Quaternary glaciations are also responsible for much of the sediments and topography along rock-avalanche travel paths in high mountains, with the kinds of consequences in the run-out zone described above.

Finally, more rock avalanches may originate in glacierized high mountain valleys than in other areas. At least this seems to be the case in the North American Cordillera, the European Alps, Karakoram Himalaya, and Southern Alps of New Zealand (Post, 1967; Evans and Clague, 1988; Hewitt, 1988, in press; McSaveney, 2002). But the number of events and, hence, their importance are likely to be greatly underestimated. Most rock avalanches occur in rugged terrain, difficult of access. Many are triggered by earthquakes or severe weather. Those that land on glaciers may be quickly buried by snow or dispersed in a few years or decades by glacier movement and ablation, further reducing the chance they will be recognized (McSaveney, 2002, p. 68). The possible scope for misidentification became evident in the 1964 Alaska earthquake, with 79 rock avalanches falling onto glaciers (Post, 1967; Marangunic and Bull, 1968; McSaveney, 1978).

Of course, large rock avalanches do occur outside the glaciated zone. The examples at the margins of active fault-block mountains in the arid U.S. southwest have been mentioned. Others are known along rugged, high-relief coastlines and in deep river gorges in mountains that have been substantially elevated in the recent geologic past. Gorges cut through different parts of the extra-glacial Himalayas, for example, host many rock avalanches (Meng et al., 2006).

8. Landscape influences of rock avalanches

Large landslides are more than just rare, isolated events, involving localized disturbance of rock walls, or ephemeral dams and additions to sediment loads. Two landscape elements endure long after a rock avalanche has occurred — the detachment zone scar and the landslide deposit. Both can influence subsequent landform development. The former may be a source of instability, rockfall, and debris flows for years or decades after the rock avalanche, and the rock wall geometry created by the landslide may persist for tens of millennia. Large landslides create constructional landforms that blanket the landscape over many square kilometers, are resistant to erosion, and can also persist for millennia.

In mountain ranges where large landslides are common, their scars and deposits exert major influences on landscape history. Moreover, recent investigations suggest that both elements affect the geomorphic development of mountain slopes and valleys in ways that extend the usual scope of mass movement studies. Many rock avalanches involve failure of a relatively
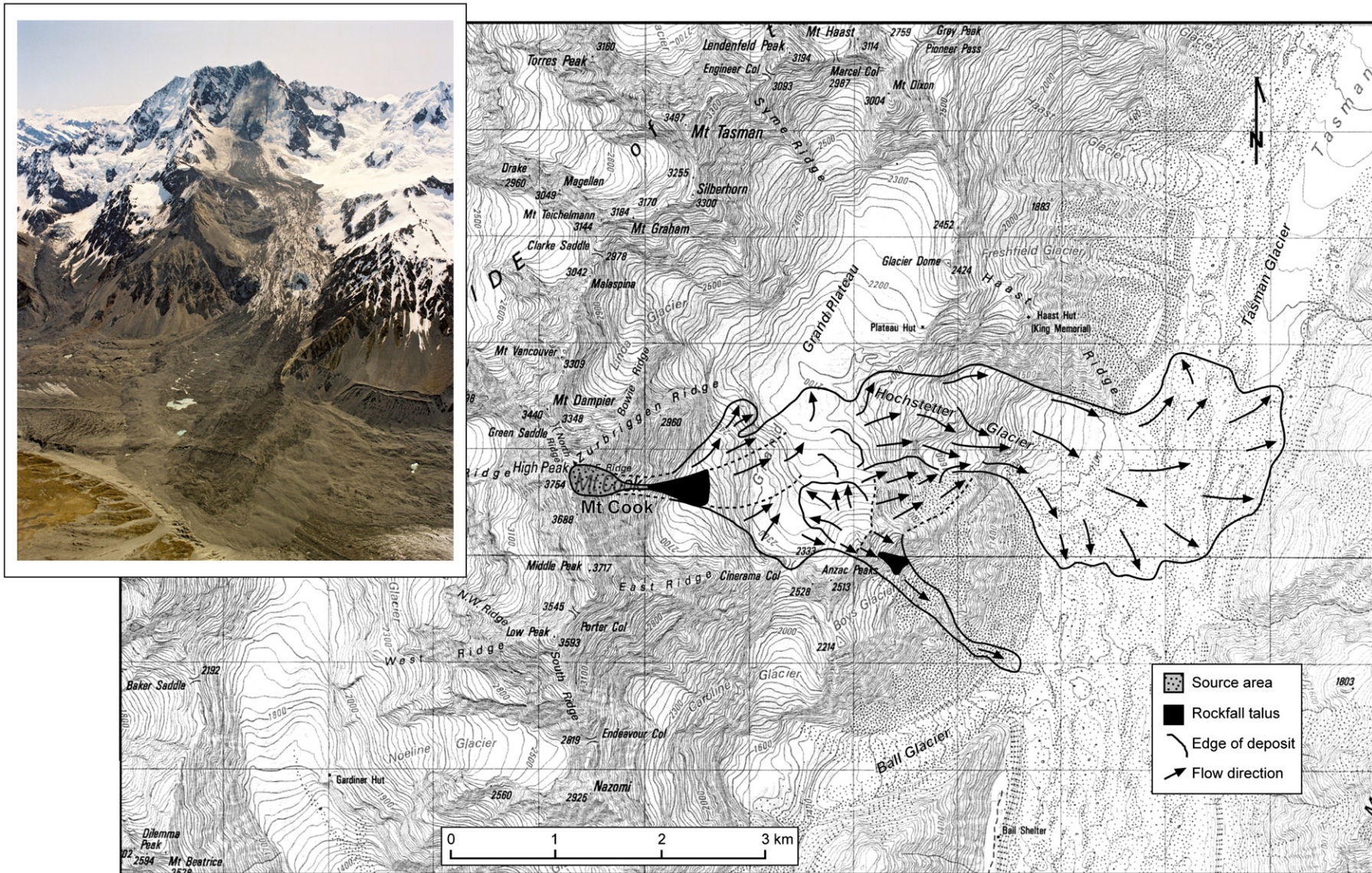
Fig. 21. Aoraki/Mt. Cook rock avalanche of December 14, 1991. The failure lowered the former summit by 10 m and shifted the highest portion 14 m to the southwest (McSaveney, 2002).
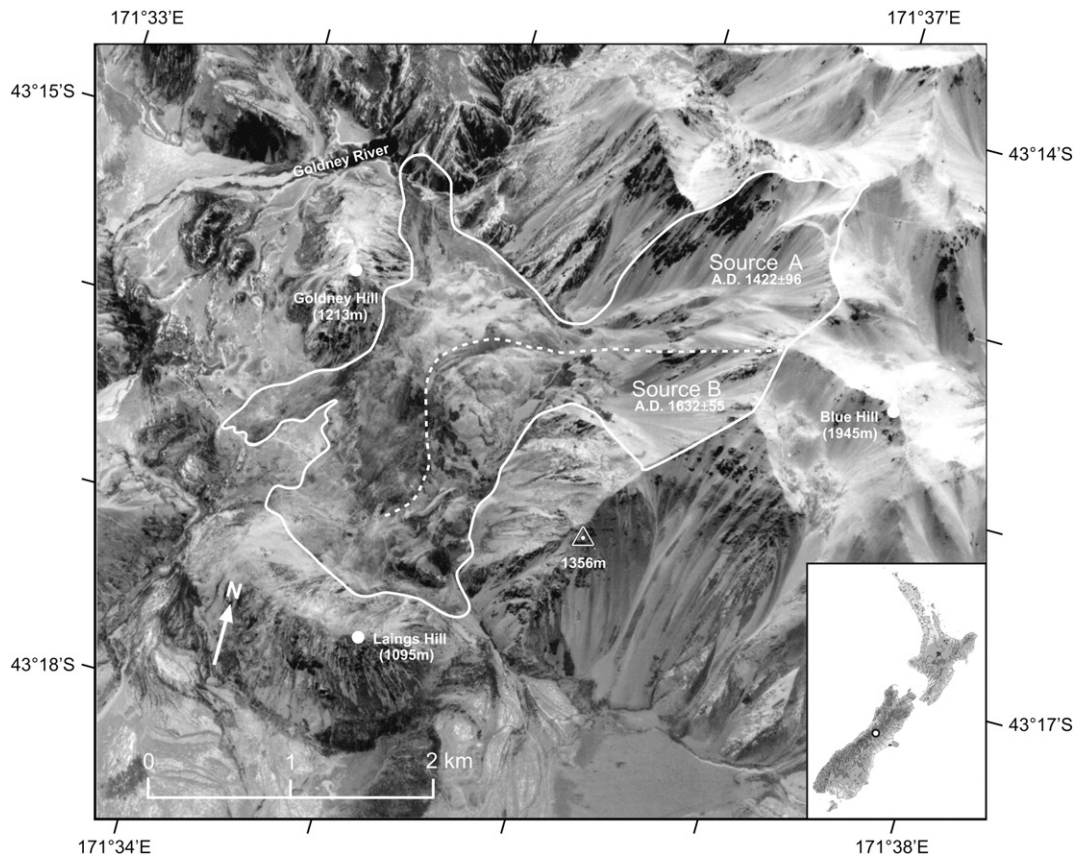
Fig. 22. Craigieburn Range rock avalanches, Southern Alps, New Zealand. Two failures occurred at this site, one in the fifteenth century and another in the late sixteenth or early seventeenth century (Orwin, 1998). The two rock avalanches significantly altered the geomorphology of the source peak and beheaded several cirques on the other side of the range. It also impacted the fluvial geomorphology of the area. (New Zealand Aerial Mapping photo no. 3B S74/1/C, Crown Copyright reserved.)

thin slab of rock, for example those at Aoraki/Mount Cook in 1991 (Fig. 21) and Mount Adams in 1999 (Hancox et al., 2005). They may shift ridge divides but do not significantly alter the form of the landscape. At the other end of the spectrum are rock avalanches arising from deep-seated failure of a thick mass of rock. They may produce cirque-like landforms that persist in the landscape (Turnbull and Davies 2006), and may shift and lower summit ridges. New Zealand examples include the historic rock avalanche at Falling Mountain (McSaveney et al., 2000) and the prehistoric events in the Craigieburn Range (Fig. 22; Whitehouse, 1981; Orwin, 1998).

Many slope process studies have focused on what happens to surfaces that connect a drainage divide or the head of slope to its base. They commonly attribute changes in slope geometry or the importance of different geomorphic processes to geotectonic conditions or climate. Broader landscape models tend to assume that the set of processes operating on slopes operate towards a dynamic equilibrium, with material removed by axial drainage. In fluvial landscapes, except for local, short-lived, and extreme cases, slope processes are thought to respond to, but not exercise control over, longitudinal or thalweg development and stream form. Stream planform and slope geometry have been attributed to a combination of climate, geological structure, tectonics, and the self-organizing evolution of the drainage

network. This paradigm has been applied both to high mountains (Scheidegger, 1991; Burbank et al., 1996; Hovius et al., 2004) and lower relief terrain (Holmes, 1978, chapter 17). Sackung and related features show that mountain ridges are by no means static (Evans, 1987; Bovis and Evans, 1996), but no one has suggested that they influence drainage basin evolution. A related, older notion interpreted dendritic drainage patterns as the normal development of stream systems, and trellised or rectangular ones as having developed under strong structural control where slopes respond accordingly (Holmes, 1978, chapter 19).

The landslides and mountain regions described here suggest a need to modify this view. The direct landscape impacts of catastrophic rockslides and rock avalanches are not confined to the space between a pre-existing interfluve and a stream channel or glacier. Their morphogenetic role is not simply to adjust slope profiles or geometry between the head and base. They involve slope modifications distinctive of the mass movement process, and their influence on stream development reflects the frequency, scale, and geometry of the landslides.

First, bedrock failures are more than immediate and powerful agents of denudation. The break-out zones in many cases extend through former ridge crests or peaks and thus change the height and position of interfluves. Second, as described above, most

of these events involve emplacement of large masses of frag-mented rock in, across, and down valleys. Their process domain, therefore, extends far beyond the source slopes to act on the axial drainage system, in some cases tens of kilometres downvalley.

In the three mountain regions of interest, rock avalanches have been widespread and recurrent throughout the Quaternary and back into the Tertiary. Their possible significance for mountain stream development and mountain morphogenesis is another "emerging issue" that interests us.

8.1. Detachment zones and watershed development

Catastrophic landslides may leave behind huge scars on rock walls and can drastically alter slope geometry. Where there are simple structural controls, as with slab failures, a geometric change in slope elements may not be evident. In many other cases, however, the slope acquires a new and different shape. A further consequence of many large rock slope failures is that they change the geometry and location of interfluves. Loss of height or of mass at higher elevations is a measure of de-nudation and change in available slope energy. Shifts in the position and height of interfluves raise questions about the evolution of mountain valley systems.

Interfluve changes due to catastrophic slope failure may happen in one or a combination of three ways:

1) *Edifice collapse*, involving partial or complete destruction of a peak or rock tower.
2) *Interfluve displacement*, where the plane of failure daylights on the reverse slope of the ridge crest.
3) *Spur collapse*, where the culmination of a buttressing ridge, or a salient on it, fails.

All three involve displacement of material or terrain that stood above a level extrapolated from the base of the source slope to the post-landslide head of the detachment zone.

Where material transfers are large, especially in edifice collapses, the mass and energy involved in the rock avalanche must be considered in reconstructing the pre-event topography, especially in reconstructions using the 'H' of Heim (1932) and in assuming the highest point of the detachment zone is the head of the post-landslide scar. In a growing number of cases, substantial excess, or "missing" mass is being invoked. This issue complicates analysis of the "rockslide" phase when, in very high mountains at least, a large vertical component of motion is important. Account must be taken of the enormous increase in crushing and pulverizing forces in the first few seconds or tens of seconds of failure before the disintegrating rock mass enters the full-blown rock-avalanche phase.

Edifice collapse seems to overlap what has been called "toppling failure" (Evans, 1987; Selby, 1993, chapter 15), but the latter suggests rotation outward from the source. "Collapse" seems a better term in many cases. McSaveney (2002) described the 1991 Aoraki/Mount Cook rock avalanche as originating with a "major collapse," taking with it "… the former summit of High Peak and a 700-m rock buttress that formerly supported it (p. 41)" (Fig. 21). Similarly, collapse of the summit and west flank of a former peak at the southern end of the Craigieburn Range, 200 km northeast of Aoraki/Mount Cook, generated two large rock avalanches about 600 and 250 years ago (Whitehouse, 1981; Orwin, 1998) (Fig. 22). A similar collapse initiated the prehistoric Gannissh Chissh event in the Kar-akoram (Fig. 23). The great landslide disaster in the Cordillera Blanca of Peru in 1970 began with collapse of the west summit and face of Nevados Huascaran (Pflaker and Ericksen, 1978,
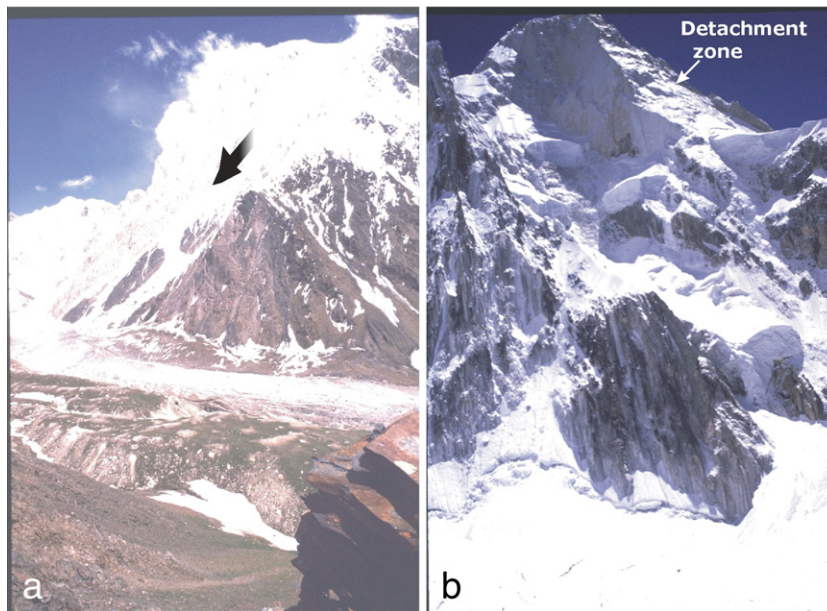

Fig. 23. Edifice collapse at Gannissh Chissh (7027 m asl) in the Karakoram. (a) The north flank and crest of the mountain collapsed (arrow) to form a rock avalanche that descended onto the glacier in the middle ground. Five to ten meters of debris were deposited on lateral moraines in the foreground. The bulk of the debris, however, traveled to the right, about 11 km down Barpu Glacier. (b) The detachment zone in sunlight, dipping 65–70°. The landslide split into two parts on this slope; one lobe surged to the site where the photo was taken, and the other flowed towards the camera station in (a).

p. 285, their Fig. 6). And the 1964 Sherman Glacier rock avalanche stemmed from the collapse of a portion of Shattered Peak (McSaveney, 1978, p. 209, his Fig. 4). Edifice collapses are common on the slopes of stratovolcanoes, both during and independent of eruptions. Collapse of the north flank of Mount St. Helens initiated the cataclysmic eruption of May 1980 (Glicken, 1990). Some peaks are sites of exceptional structural or rock strength, but in active orogens, including those discussed here, most peaks comprise damaged and defective rock of low in situ strength. On time scales of 10^4–10^6 years, the array of mountain peaks may change in position and height due to retreat of interfluves associated with repeated edifice collapse. McSaveney (2002) suggests that in the Aoraki/Mount Cook region of the Southern Alps, all peaks are at or very close to imminent collapse and that minor mass wasting events can trigger large catastrophic failures. In many cases, the structural integrity of peaks may be more related to protective ice carapaces or permafrost than rock strength. Most of the mountain peaks and ridges in the three regions described here are draped with snow and ice, the state of tectonically damaged rock hidden and unknown under glaciers, cornices, and perennial snow.

Interfluve displacement lowers a watershed and shifts its position relative to intervening valley floors. Edifice or spur collapse may also be involved, but is difficult to document unless pre-existing topography is known. The Köfels event in the Alps considerably lowered the ridge above the source slope, shifting the interfluve westward (Heuberger, 1975; Heuberger et al., 1984). The pre-event topography must be considered in reconstructing the energy needed for the enormous climb and transport of material up the opposing, east slope to the Tauferberg and beyond. A reconstruction of one of the largest known terrestrial rockslide — rock avalanches, the prehistoric Saidmarreh event in the Zagros Mountains of Iran, shows major lowering and displacement of the Kabir Kuh interfluve at the source (Harrison and Falcon, 1938). Similar changes apply in the giant Flims event in Switzerland (von Poschinger, 2002, p. 247, his Fig. 11; von Poschinger, 2004) and the Craigieburn rock avalanches in New Zealand (Whitehouse, 1981; Orwin, 1998). Interfluve changes also occurred during the prehistoric Villavil events in Argentina (Fauque and Tchilinguirian, 2002, p. 311, their Fig. 8), the 1903 Frank, Alberta event (Cruden and Krahn, 1978, their Figs. 5–8), and the 1958 Madison Canyon event, Montana (Hadley, 1964).

Spur collapse is a third mechanism for changing local watershed geometry or relief. Continued rock-avalanche failures from Mount Fletcher in the Southern Alps illustrate the modification of a spur and the shift of an interfluve between tributary hanging valleys in glaciated terrain. The Fernpass event in the Austrian Alps is another example (Abele, 1974).

Interfluve displacement by rock avalanches may be a common high-elevation response to valley incision. McSaveney (2002, p. 68) suggests that "…high on mountain slopes… gravitational collapse is the only operative process to keep pace with glacial and fluvial incision…" He illustrates this point with several New Zealand examples and suggests that it is not necessary to invoke some unique predesign or strength of peaks. He does, however, suggest that zones of exceptional uplift and faulting, like the Main Divide fault in the Southern Alps, have higher incidences of collapse than more passive mountains. The same can be said for the active Raikhot and Stak fault systems in the Karakoram, where there are exceptional numbers of large rock avalanches (Shroder et al., 1989; Shroder, 1993). It is also one of the most rapidly uplifting regions in the entire Himalaya and the site of some of the greatest relief and steepest terrain on Earth (Searle, 1991, p 321). Having said this, most of the numerous examples in the upper Indus Basin are not associated with any known active faults, and some of the largest rock avalanches in this region have descended from slopes with 1500–2000 m of relief, much less than the maximum local relief of 5000–7000 m (Hewitt, 2001, 2006b).

In any case, such events impact the morphological evolution of mountain valleys. They not only contribute to net denudation but also change valley spacing and form by shifting interfluves. Individual events may seem small — a few tens to hundreds of meters of vertical and horizontal change, perhaps over watershed lengths of 2 to 3 km. However, high numbers and densities of Quaternary landslides in the three regions emphasized here suggest a contribution to watershed change that may be comparable to changes effected by rivers.

In the Karakoram Himalaya, more than 270 rock avalanches larger than 10^6 m^3 have been identified within about 20% of the region. This number almost certainly is an underestimate of the spatial frequency of large catastrophic rock slope failures in the region, because the frequency is much higher in the glacial zone where most deposits are removed in a few decades and detachment zones are masked by snow and ice (Hewitt, 2005). However, if existing data are at all representative, they translate into at least 1000 events in the 100,000 km^2 area of the Upper Indus Basin. Nearly all of these events are Holocene in age, suggesting an incidence of at least ten per century, similar to estimates for the Southern Alps (McSaveney, 2002). Only five of them are known to have occurred in the past 100 years — or seven counting the three separate events at Bualtar in 1986 (Hewitt, 1988). Only three historic examples are known to be outside the glaciated zone, all dating to the nineteenth century. It is likely more events were missed than reported, but even if these estimates are of the right order, rock avalanches must play a significant role in reorganizing fluvial systems and dictating geomorphic evolution in the Karakoram.

A more significant statistic may be the average spacing of rock avalanches along stream valleys. Hewitt (2004) sampled a 1000-km length of the Upper Indus streams and found one rock avalanche per 14 km of river length. Similar or greater densities are reported from parts of the Andes (Fauque and Tchilinguirian, 2002; Hermanns et al., 2006), and Norway (Blikra et al., 2006). Most of the rock avalanches that have been identified are interfluve-modifying events. The implications for stream and ridge spacing are intriguing, because the geomorphic changes extend from the heads of slopes to their bases.

It also is important that the changes are irreversible and occur in bedrock. Much of the erosional action of streams is expended on sediment, is reversed in aggradation cycles, and, as shown below, tends to be reversed or stalled by landslide interference. Fluvial responses to interfluve changes caused by catastrophic failures are complicated by the long runout of rock avalanches in valleys.

8.2. Landslide-interrupted valleys

Landslide barriers strongly influence valley-floor morphology and sediment transport and deposition in mountains. Differences in channel morphology downstream, within, and upstream of rock-avalanche dams have been summarized by Whitehouse (1983, p. 271), based on observations in the central Southern Alps: "...River gradients are commonly low above the [rock-avalanche] deposit with marked aggradation, while deeply incised gorges may occur below or through the deposit."

An alternation of steeper, narrower stream sections and gentler, more open sections is characteristic of major rivers and their tributaries in all three regions of interest and many others (Hewitt, 2006a).

Landslide barriers can perturb a drainage system for tens of thousands of years, in some cases irreversibly. They alter the entire river system by regulating the throughput of sediment and water in and below the affected reach (Fig. 24). The dimensions and shape of the impounded reach are important,

but the effects of the landslide barrier on base level and river planform extend much farther downstream and upstream. The barriers may also shift drainage divides. Two prehistoric rock avalanches in northern South Island of New Zealand pond Lake McRay between them on a former interfluve (M. McSaveney, written communication, 2007). And the barrier emplaced by 1965 Hope landslide in British Columbia forms the influve between northwest-flowing Nicolum River and southeast-flowing Skagit River (Mathews and McTaggart, 1978).

Many cross-valley rock avalanches impound lakes. The stability of the dams, the possibility of catastrophic failure, and outburst events have received much attention (Costa and Schuster, 1988; Clague and Evans, 1994a; Schuster, 2000) and are obviously important for hazard assessment. Many large landslide dams persist for centuries and, in partially breached states, for thousands of years. Many of the dams that have been identified in the three regions of interest were infilled or drained long ago. They are known mainly from remnants of lacustrine sediments deposited behind former dams. In the Upper Indus basin, few large landslide-dammed lakes exist today, yet more than 160 former ones have been found, some over 100 km long and 500–1000 m deep at the barriers (Hewitt, 1998, 2001, 2004).

As long as a landslide barrier remains intact, the valley immediately upstream is the site of basin-wide sedimentation. In narrow mountain valleys, the basin created by the barrier may
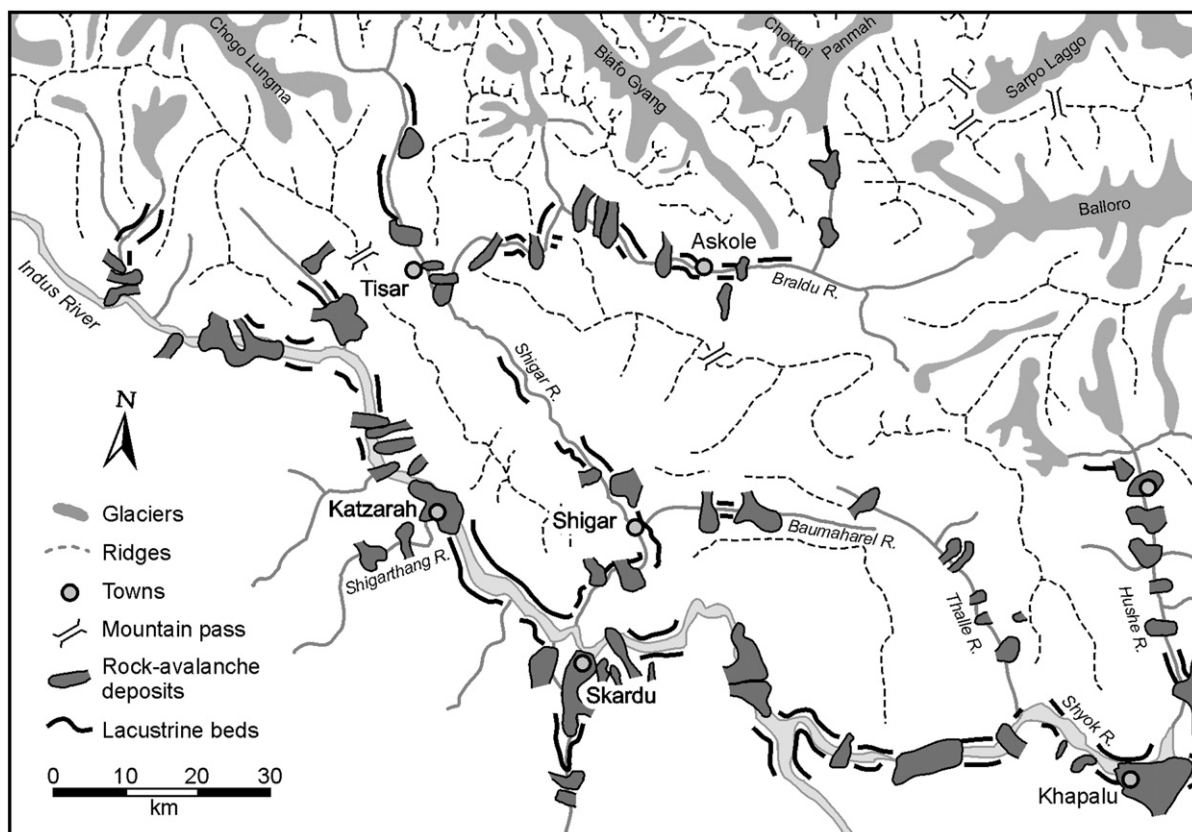

Fig. 24. Rock avalanches that have dammed Upper Indus streams in Baltistan (Hewitt, 1998).

be small and thus fill rapidly. Conversely, large volumes of fine sediment may accumulate in broader, lower-gradient valleys over periods of millennia, even millions of years. In either case, abundant coarse sediment is stored in deltas and below active slopes, and water leaving large impoundments may carry little or no sediment.

In quiet reaches of lakes below stable slopes, fine-grained lacustrine sediments fill the valley, building rhythmic sequences that are thick and extensive. Coarse sediments from tributary catchments feed rapidly advancing deltas that build wedges of sand and gravel into the lake, interfingering with other lake sediments. Eventually, braided or anastomosing rivers deposit a cap of coarse sediment over the lacustrine fill. This classic sequence requires that the dam survive long enough for the reservoir to fill with sediment. More than a dozen, intact rock-avalanche dams exist in the Karakoram Himalaya, with complete fill sequences upvalley of barriers, and many prehistoric dams survived until filling was complete. Infilling can also occur following a partial breach, as has been described at Köfels and Flims (von Poschinger, 2004).

Eventually, a complex sedimentary assemblage develops, characterized by overlap and interfingering of diverse lithofacies. The basin fill may provide a record of individual depositional events and seasonal and longer-term sedimentation episodes. Some valley fills resulting from aggradation episodes are vast. The floors of the Skardu-Shigar basins in the Karakoram, which have a total area of about 470 km^2, are underlain by thick sediments deposited behind the Katzarah rock-avalanche barrier. Sixteen large sediment fans extend onto the floor of the basins. Deposition against the original barrier extended to 70 m above present river level, and the fill behind the existing, partially breached barrier is more than 100 m thick (Hewitt, 1998). The shape of the basin floor before the Katzarah rock avalanche is unknown, but assuming the sediment wedge decreases in thickness to the head of the Skardu and Shigar basins, it has a volume of about 55 km^3. Some of the fill has been removed by river incision, but about 35 km^3 remains. Other huge volumes of sediments impounded by rock avalanches exist along the Shyok, Gilgit, Hunza, Chapursan, Ishkoman, Darkot, and Upper Chitral Rivers (Hewitt, 2001, 2002a). Satellite images taken at times of high summer flows give the impression that many of the barriers still impound lakes (Searle, 1991, his Fig. 12).

Large lakes dammed by rock-avalanche deposits in other mountain ranges include Lake Waikaremoana (56 km^2) in New Zealand, which formed about 2200 years ago (Korup, 2002), and Lake Sarez in the Pamirs of Tajikistan, a 53-km-long body of water impounded in 1911 (International Strategy for Disaster Reduction, 2000). Dozens of smaller examples are known (Hadley, 1964; Eisbacher and Clague, 1984; Schuster, 1986; Abdrakhmatov et al., 2004).

Downstream of the landslide dam, as with any impoundment, the sediment-starved main stream may incise the valley fill or bedrock (Collier et al., 1996). Associated knick points may move into tributary valleys. Conversely, in high mountain valleys, tributaries may build coarse cross-valley fans that an unimpeded main stream would remove. In such a situation, aggradation may occur downstream of the barrier as well as upstream, a common occurrence, for example, along Upper Indus streams (Hughes, 1984).

Fig. 25. Part of the Canterbury Plains with the Southern Alps in distance. The plain is underlain by thick non-glacial and outwash gravels shed from the Southern Alps. Some of the gravels may be derived from large landslides in these mountains. (GNS Science photo no. 6883-13).

8.3. General consequences of landslide-fragmented rivers

By impounding, obstructing, or diverting rivers, landslides disrupt stream continuity and mask their responses to climate change. The systematic evolution of the geomorphic system, which is generally envisaged to be a response to climatic or tectonic forcing, is also disrupted. The river directly downstream of impounded reaches may bear little or no relation to what would be expected based on stream order, local relief, climate, hydrology, vegetation, and geology. Put another way, these variables do not explain differences in and between valleys over the course of the landslide interruption cycle.

Stream behavior is chronically disturbed by sediment storage behind a landslide barrier and by subsequent mobilization of the sediment as the barrier degrades (Hewitt, 2002a, 2005; Korup et al., 2004, 2006). Aggradation or incision can occur without other changes in the regional or basin environment. Interruption episodes may overlap and interact in basins with many large slope failures and barriers. An additional complication is that disruption episodes are rarely in phase with one another. Interruptions may occur more-or-less independently and differ widely in scale and scope, but their outcomes are not independent. Multiple diachronous cycles of aggradation, degradation, and superposition are reflected in valley landforms. The overall result is a naturally fragmented drainage system in which fluvial landforms and valley fills are controlled by the pattern and sequence of landslide interruptions.

Fragmented river systems are a widely recognized result of human intervention and control works (Dynesius and Nilsson, 1994). In contrast, natural interference has been seen as accidental and of minor importance, even though many rivers in high mountains display this phenomenon.

8.4. Extra-montane impacts

Large landslides can affect landscapes outside the mountain ranges in which they occur. The principal effect is one of facilitating the transfer of sediment from rising mountains to lowlands, coastal plains, and oceans. Landslides of all sizes have long been recognized as contributing to denudation of mountains, but mass movements are complexly and intimately coupled to fluvial and glacial systems.

The impact of mass movements on extra-montane settings is clearly illustrated on the South Island of New Zealand. The Southern Alps, a relatively narrow, but rapidly rising mountain range, is flanked by low, coalescent outwash fans of Pleistocene and late Tertiary age (Fig. 25). The Canterbury Plains to the east of the Alps are about 70 km wide and 240 km long; much of them are underlain by gravelly outwash deposited during the last glaciation. Inset into the outwash plain are the broad, Holocene braidplains of rivers flowing from the Southern Alps to the Pacific Ocean, which attest to continuing high rates of sediment transfer from the mountains. What proportion of the sediments of the Canterbury Plains derives indirectly from landslides is unknown, but it is undoubtedly significant. To the west of the Alps and bordering the range along the Alpine Fault

is a late Tertiary and Quaternary coastal lowland ca. 40 km wide and 400 km long. Large areas of this lowland are underlain by sediments deposited in the last few thousand years.

Prehistoric earthquakes on the Alpine Fault probably triggered landslides into mountain valleys. The blockages likely perturbed the fluvial system, leading to large-scale transfers of sediments onto the coastal plain (Wells and Goff, 2007). Earthquakes, however, are not required to explain these large sediment transfers. Storms producing daily rainfall in excess of 1 m are not unusual in the Southern Alps and can trigger debris and rock avalanches that lead to aggradation of lowland floodplains (Griffiths and McSaveney, 1983, 1986; Hicks et al., 1990). Similar transfers of huge volumes of sediment characterize areas flanking uplifting mountains in Alaska, Taiwan, New Guinea, and the Himalaya.

9. Concluding remarks

The emerging issues discussed in this paper point to a changing emphasis in research on catastrophic rock slope failures. To identify the implications of this shift, we return to views of the field mapped out in its literature.

In many ways, the prevailing approach over the past half century was established, at least for the English-speaking world, in the two benchmark volumes of Voight (1978).

Emphasis was on integrating geoscience and engineering to gain a better understanding of landslide events, triggers, and processes. Studies drew on rock mechanics, geomorphology, and numerical modeling to explain the high velocity and excess runout of rock avalanches. The important sub-field of research on large landslide dams has employed similar methods, while extending them to investigate dam stability, causes of catastrophic failure, and the behavior of outburst floods (Schuster, 1986; Costa and Schuster, 1988).

It would be difficult to overemphasize the importance and contribution of this approach. Most of the emerging issues identified here have arisen from studies done in that vein. However, as noted at the beginning of the paper, most published studies of rock avalanches examine single events. They may seek to identify more-or-less universal governing properties, but present a portrait of the singular, extreme event in isolation. Out of this has come a style of analysis resembling the 'accidental' geomorphology of earlier times (Cotton, 1958) and a tendency to perpetuate a catastrophist view of events (Scheidegger, 1975). Geological, geotechnical, and modeling concerns have developed a focus wholly on the landslide itself or on geologic or meteorological controls that bear solely on it. Some comparative studies have drawn on examples from many different parts of the world, but essentially as isolated and independent points in data sets defined by abstract parameters for the characteristic or ideal landslide, in the search for global generalizations (Keefer, 1984; Evans, 2006).

Again, we emphasize that the benefits of this approach are considerable. However, they have come at the expense of focused studies of regionally significant conditions and risk assessments and, especially, relations of landslides to other surface processes, the roles of landslides in mountain morphogenesis and denudation, and their place in broader Quaternary Earth history. These

are the concerns emphasized here. They depend on landslide science, but require recognition of other approaches and other geoscience and environmental issues.

What seems to be happening now is a convergence of several, formerly separate approaches amid growing evidence that catastrophic landslides are more common than previously thought. Some of the more evident developments arise from questions identified with:

1) Tectonic geomorphology and related denudation and sediment delivery in mountain watersheds:
   Regional tectonics, seismicity, geological structure, lithology, climate, and relief, are now being seen to act together in preconditioning rock slopes for failure and in determining the incidence and behavior of large landslides. Regional investigations of sets of landslides are revealing previously unrecognized aspects of their incidence, temporal and spatial patterns, and diversity (Hermanns and Strecker, 1999; Hewitt, 2002a; Jackson, 2002; Abdrakhmatov and Strom, 2006; Hermanns et al., 2006). At the same time, studies of the role of landslides in regional denudation are revealing links to tectonic, climatic, and drainage-system dynamics (Burbank et al., 1996; Whipple, 2001; Korup et al., 2006; Hovius and Stark, 2006).

2) Landslides in glacial environments and late Quaternary history:
   In the three orogens emphasized here, glaciation has had a major influence on the style and incidence of late Pleistocene and Holocene rock slope failures (Eisbacher, 1979; Bovis and Stewart, 1998; Blikra et al., 2006). Rock avalanches not only are affected by runout over snow and ice, but also influence glacier activity (Tarr and Martin, 1912; Post, 1967; Evans and Clague, 1988; Hewitt, 1988, in press; Deline, 2005). The intimate relation between glaciers and rock avalanches is an emerging concern in glacial geomorphology (Kirkbride, 1995), involving problems in identifying and assessing contributing processes and depositional legacies.

3) Comparative diagnostics and deposit relations:
   Criteria for recognizing rock-avalanche deposits, described above, are part of several, on-going and converging lines of enquiry. First, although some attributes of rock-avalanche deposits are singular and definitive, notably those relating to lithology and processes of rock fragmentation, deposit morphology, facies properties, and sedimentary architecture can differ considerably. A variety of distinct, but related forms arise from complex runout and emplacement, or from stalling at intermediate stages, due to interactions with terrain and substrate (Strom, 1998; Hewitt, 2002b). Second, efforts are being made to define reliable field or remotely sensed diagnostics for sediments that have been, or may be, mistaken for rock-avalanche materials. Such materials have been variously described as coarse, angular, immature sediments, breccias, diamicts, or "fragmentites" (Laznicka, 1988), and arise from several types of mass movement and glacial processes, from catastrophic floods, and from tectonic cataclasis. Similar materials of different origin can be closely associated and even interbedded in mountain valleys.

Third, additional problems arise from interactions among different Earth surface processes. Important cases discussed above are entrainment of rock-avalanche material by glaciers and rivers, and deposition of rock-avalanche material in lakes and coastal waters. These interactions can generate deposits that are intermediate between the primary rock-avalanche material and glacial, fluvial, or lacustrine sediments.

It may not be possible to make an adequate diagnosis based solely on exposed sediments and samples. For efficiency and greater confidence, the researcher must identify and reconstruct broader landscape developments, form-and-process relations, and associations of landforms and deposits. These larger-scale issues lead into and enhance the value of:

4) A regional landscape or land-system perspective:
   We argue that the significance of catastrophic rock slope failures lies in interactions of the landslide process with glaciers, streams, and the coastal zone, perhaps even more than in the mass movement event itself (Hewitt, 1998; Fort, 2000; Hermanns et al., 2006; Korup et al., 2006). As a result, greater attention is drawn to how landslides relate to regional environments and distinctive mixes of Earth surface processes. Geomorphic and stratigraphic interpretations of these phenomena require an approach that can incorporate and relate the range of process, landforms, and sediments in given landscapes. One such approach is that of the "land system," pioneered by Eyles (1983).
   Consider, for example, the complex of landforms and sediments associated with landslide dams and their subsequent reworking during barrier breaching. Although each element of this complex is distinct, they collectively represent a landform-sediment association comparable to, for example, the "glaciated valley landsystem" of Eyles (1983, pp. 91–110). Individual examples are of comparable dimensions — kilometers across and tens of kilometers along, and valley fills tens to hundreds of meters thick. The phases described above can be used to define the main features of a "landslide-interrupted valley land system" (Hewitt, 2006a).

If these developments suggest a shift of emphasis, they also invite us to look to an earlier literature that shows an awareness of broader concerns. Notable is the continuing influence of the overview of Heim (1932) on Alpine landslides. He looked at regional conditions to explain a large set of events, emphasized the particulars of each event and its setting, and classified different forms of rock slope failure and resulting mass movements. Abele (1974) continued his work, but with results not widely appreciated until after his death (von Poschinger, 2002). Some older geomorphic investigations that focused on the landscape recognized phenomena that have tended to disappear from the literature. Their concerns and terminology have been revisited, for example, in the recognition of "epigenetic" or superimposed gorges, "barrier-defended" terraces, and other features associated with landslide-interrupted mountain streams (Cotton, 1958; Hewitt, 2006a).

In general, these developments suggest a reversal of the tendency to pursue a landslide science that is independent of other geoscience, geomorphic, environmental, and geohazard issues.

They lead to a greater emphasis on suites of landslides, regional contexts, and landscape genesis. What may appear as an undue emphasis on diversity, complexity, regional contexts, and sub-field integration arises more in relation to the prevailing approach than as something advocated as intrinsically desirable or necessary. The interaction of large landslides with other processes, and with terrain, introduces complexities and heterogeneity, and it requires new initiatives at the interfaces between specialized fields. It does not mean that other and different ways to simplify and model seemingly complex situations, or other strategies of generalization, will not arise. Some researchers have already begun to explore ways to address these questions without being buried in endless detail and diversity (Hovius et al., 2000; Hewitt, 2005; Korup et al., 2006).

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