The Varnes classification of landslide types, an update

Article *in* Landslides · April 2014

DOI: 10.1007/s10346-013-0436-y

3 authors:

Oldrich Hungr
University of British Columbia - Vancouver

Serge Leroueil
Laval University

Luciano Picarelli
Università degli Studi della Campania "Luigi Vanvitelli

Landslides
DOI 10.1007/s10346-013-0436-y
Received: 22 April 2013
Accepted: 23 September 2013

Oldrich Hungr · Serge Leroueil · Luciano Picarelli

The Varnes classification of landslide types, an update

Abstract The goal of this article is to revise several aspects of the well-known classification of landslides, developed by Varnes (1978). The primary recommendation is to modify the definition of landslide-forming materials, to provide compatibility with accepted geotechnical and geological terminology of rocks and soils. Other, less important modifications of the classification system are suggested, resulting from recent developments of the landslide science. The modified Varnes classification of landslides has 32 landslide types, each of which is backed by a formal definition. The definitions should facilitate backward compatibility of the system as well as possible translation to other languages. Complex landslides are not included as a separate category type, but composite types can be constructed by the user of the classification by combining two or more type names, if advantageous.

Keywords Classification of landslides · Typology · Materials · Mechanisms · Engineering geology · Geotechnical engineering

Introduction

The system of landslide classification devised by the late D.J. Varnes has become the most widely used system in the English language (Varnes 1954, 1978; Cruden and Varnes 1996). Its sustained popularity in North America and its variations in all other continents attest to its usefulness. The authors do not intend to propose an entirely new landslide classification system but aim to introduce modifications to the Varnes classification to reflect recent advances in understanding of landslide phenomena and the materials and mechanisms involved. The starting point of the modifications is the 1978 version of the classification (Varnes 1978), taking also into account concepts introduced by Cruden and Varnes (1996).

Type of material is one of the most important factors influencing the behavior of landslides. However, the threefold material division proposed by Varnes (1978), including "rock, debris, and earth," is compatible neither with geological terminology of materials distinguished by origin, nor with geotechnical classifications based on mechanical properties (e.g., Morgenstern 1992; Leroueil et al. 1996). Thus, characterization of materials appears to be one aspect of Varnes' classification that warrants updating. In addition to this important change, several other changes, related primarily to movement mechanisms, are described below.

Focus of the classification

A landslide is a physical system that develops in time through several stages (e.g., Terzaghi 1950; Leroueil et al. 1996). As reviewed by Skempton and Hutchinson (1969), the history of a mass movement comprises pre-failure deformations, failure itself and post-failure displacements. Many landslides exhibit a number of movement episodes, separated by long or short periods of relative quiescence. The following definition of the term "failure," inspired by a discussion by Leroueil et al. (1996) is proposed for the purposes of this paper:

Failure is the single most significant movement episode in the known or anticipated history of a landslide, which usually involves the first formation of a fully developed rupture surface as a displacement or strain discontinuity (discrete or distributed in a zone of finite thickness, cf. Morgenstern and Tschalenko 1967).

The degree of strength loss during failure determines the post-failure velocity of the landslide. The failure stage may involve a kinematic change from sliding to flow or fall, which is also relevant to post-failure behavior and destructiveness of the landslide.

Cruden and Varnes (1996) proposed separate names for the movement mode during each stage of a given landslide. This is a desirable goal during detailed investigation and reporting. However, for communication, we also need to be able to assign simple names to the whole landslide process and such names should be compatible with established terminology.

One practical statement illustrating the need for a typological classification was given by Professor J.N. Hutchinson (personal communication, 2000, paraphrased): "To provide labels for a filing system to store scientific paper reprints. A well-organized system will help the user to rapidly locate articles dealing with a given phenomenon and its typical characteristics." A similar system of labels is needed also in one's mind, to organize facts and ideas relevant to a given class of phenomena and communicate them to others. Of course, different individuals have different priorities and a classification system should be flexible enough to accommodate their needs.

To give an example: A landslide may begin with slow pre-failure deformation and cracking of surficial soil on a steep hillside. Then a shallow sliding failure develops. The landslide mass accelerates, disintegrates, enlarges through entrainment and becomes a flow-like debris avalanche. The avalanche enters a drainage channel, entrains water and more saturated soil and turns into a surging flow of debris. On entering a deposition fan, the flow drops the coarsest fractions and continues as a sediment-laden flood. This is a complex process. Yet, it is a common one and we should be able to apply the simple traditional term "debris flow" to the whole scenario. Otherwise, an article about such an event would need to be torn into fragments, before it can be filed. Several such comprehensive terms have been established in the professional literature for more than 100 years.

It is proposed here that the simple term assigned to a given landslide type (or a specific case) should reflect the particular focus of the researcher. If he or she is concerned with the runout of the event, then the overall term "debris flow" is appropriate. If the main focus is the pre-failure mechanism in the source area, then "debris slide" or "slope deformation" may be more relevant. The system should be flexible enough to accommodate all such uses. It should be left to the user whether he/she finds it advantageous to construct a composite class such as "translational rock slide—rock avalanche," within the framework of the classification system.

Even at a price of certain simplification, each class should be unique. A class defined as "complex" is not useful. Almost every landslide is complex to a degree. Thus, a "complex" class could hold most of the information, without the need for any other classes.

Published online: 30 November 2013

Additional objectives for a classification system

The number of classes should be reasonably small, to make the system simple and easy to use and review.

The system should be respectful of previous usage and adopt established terms to the greatest extent possible, to enhance "backward compatibility" with older literature.

The system should be sufficiently flexible to allow application both in cases where only meager preliminary data exist, as well as those where data are detailed and abundant.

Each class name should be supported by a concise, but comprehensive formal definition. Such class definition paragraphs can be translated to different languages without difficulty and class names can be attached in various languages according to established local usage. The principles of the classification will thus remain valid, repeatable, and refutable, regardless of the actual words that are used in forming the class name.

Brief history

Some of the earliest landslide classification systems originated in the Alpine countries. Baltzer (1875) in Switzerland seems to have been the first to distinguish between the various basic modes of motion: fall, slide, and flow. This division persists to the present time, supplemented by toppling and spreading (Fig. 1).

Several authors, including Heim (1932) and Zaruba and Mencl (1969) focused on landslide types that are characteristic of given material facies described in geological terms.

Debris flows represent a particularly important hazard in mountainous terrain and have attracted special attention from early days. The classic Austrian monograph "Die Muren" by Stini (1910) brings attention to the variety of debris movement in mountain channels, ranging from floods to debris-charged floods ("Muren") to boulder-fronted, surging debris flows ("Murgänge"). Similar phenomena have been described in the arid regions of the southwestern USA as "mud flows" by Bull (1964) and others. Debris-charged "hyperconcentrated" floods have been studied extensively on the volcanoes of the US North-West (e.g., Pierson 2005; Vallance 2005).

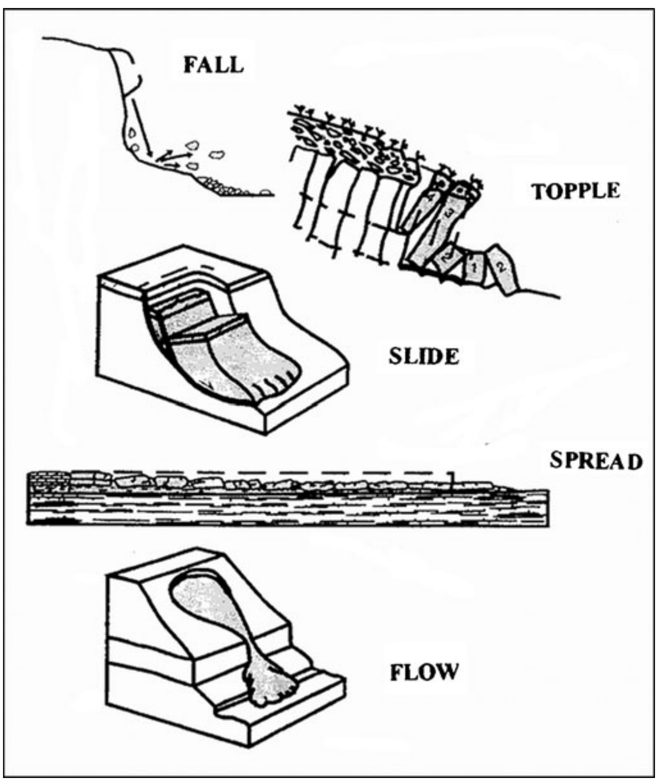

Fig. 1 Types of movement (Cruden and Varnes 1996) The scale of the diagrams could vary from a few metres to hundreds of metres as shown by examples in the paper

In the USA, Sharpe (1938) introduced a tri-dimensional classification system recognizing type of movement, material and movement velocity. He also coined (presumably) the important terms debris flow (channeled), debris avalanche (open-slope), and earth flow.

The term "earth flow" was reinforced and thoroughly described in the work of Keefer and Johnson (1983) and is used in North America as a synonym for the British "mudslide" (Hutchinson 1988). The latter word is frequently misused in media reports. Therefore, "earth flow" is preferable.

Sharpe's framework was expanded by Varnes (1954, 1978) in his influential articles prepared for the Transportation Research Board of the National Research Council in Washington. This was modified in 1996 by Cruden and Varnes, to concentrate on the type and rate of movement. The 1978 version of the "Varnes Classification System" was widely accepted by workers in many countries, albeit usually with modifications (e.g., Highland and Bobrowsky 2008; Dikau et al. 1996).

The "Varnes classification," is summarized in a poster-format Fig. 2.1 of Varnes (1978, as simplified in Table 1 in this paper). Here, within the framework of a matrix whose rows represent the type of movement and columns the type of material, are 29 landslide type names or keywords, which are further defined and described in the text of the paper. A velocity scale, later updated by International Geotechnical Society's UNESCO Working Party on World Landslide Inventory (WP/WLI) (1995) and Cruden and Varnes (1996) completes the classification (Table 2).

In England, Hutchinson (1968, 1988) developed a system without a matrix framework, utilizing multiple dimensions such as material, morphology, water content, rate, kinematics, and focusing on failure and propagation mechanisms. An attempt to correlate Hutchinson's and Varnes' systems specifically for flow-like landslides was published by Hungr et al. (2001).

Experts interested in landslide classification are most often engineering geologists. Geotechnical engineers have been concerned primarily with sliding movements and have not developed a complete set of landslide names, concentrating instead on the classification of materials. A simplified system based primarily on geotechnical concepts such as liquefaction and pre-shearing of clays was proposed by Sassa (1999).

One important engineering contribution is the term "flow slide," designating an extremely rapid failure resulting from the liquefaction of saturated sand (Casagrande 1940), or remolding of sensitive clay (Meyerhof 1957). The term has long been widely used in geotechnical practice and it has important practical implications (e.g., Terzaghi and Peck 1967).

Rock engineers contributed the terms "wedge slide" (Londe 1965; Hoek and Bray 1981), "flexural topple," and "block topple" (Goodman and Bray 1976). Specialized classifications have been devised for rock slope deformations (Hutchinson 1988), subaqueous landslides (e.g., Postma 1986), landslides in permafrost (McRoberts and Morgenstern 1974), and in sensitive ("quick") clay (Locat et al. 2011).

Table 1 A summary of Varnes' 1978 classification system (based on Varnes 1978, Fig. 2.1)

Movement type	Rock	Debris	Earth
Fall	1. Rock fall	2. Debris fall	3. Earth fall
Topple	4. Rock topple	5. Debris topple	6. Earth topple
Rotational sliding	7. Rock slump	8. Debris slump	9. Earth slump
Translational sliding	10. Block slide	11. Debris slide	12. Earth slide
Lateral spreading	13. Rock spread	–	14. Earth spread
Flow	15. Rock creep	16. Talus flow	21. Dry sand flow
		17. Debris flow	22. Wet sand flow
		18. Debris avalanche	23. Quick clay flow
		19. Solifluction	24. Earth flow
		20. Soil creep	25. Rapid earth flow
			26. Loess flow
Complex	27. Rock slide-debris avalanche	28. Cambering, valley bulging	29. Earth slump-earth flow

A Working Party of the International Geotechnical Societies, sponsored by UNESCO, produced a series of "suggested methods" for the World Landslide Inventory (International Geotechnical Society's UNESCO Working Party on World Landslide Inventory (WP/WLI 1990, 1991, 1993a, 1993b, 1994, 1995). These documents provide useful methodologies for preparing landslide reports and describing landslide causes, degree of activity and movement rate.

Landslide material terminology

Geotechnical material terminology

The authors' view is that geotechnical material terminology is most useful, as it relates best to the mechanical behavior of the landslide. To describe materials modified by geomorphic processes, including landsliding itself, it is necessary to supplement the geotechnical terms by names of mixed materials, namely "debris" and "mud," as described later in this section.

The proposed list of material types, compiled by means of a simplification of existing soil and rock description systems, is summarized in Table 3. The first column of the table lists the material types that can be used directly in forming landslide names.

These types: "rock," "clay," "mud," "silt," "sand," "gravel," "boulders," "debris," "peat," and "ice" replace the former threefold material classes used by Varnes (1978). Characteristics listed in the second column of the table can be used as supplementary terms. For example, "sensitive" clay will lose strength on remolding, while "partially saturated" silt may lose apparent cohesion on wetting.

Of course, many soils are transitional between textural classes. It is suggested that transitional terms be simplified to the component that is the most significant in terms of physical behavior. For example, a clayey silt should be called *silt* if it has very low plasticity, or *clay* if it is plastic.

Where the landslide source contains alternating zones of various materials (e.g., sand and clay), the material that plays the dominant role in the failure or propagation mechanisms should be used, even at the cost of a certain subjectivity. Where a dominant component cannot be identified, it is possible to use two terms, e.g., "rock and ice avalanche."

The words "debris" and "mud" do not have clear equivalents in geotechnical terminology, but have acquired status in geology and landslide science and have therefore been retained (Bates and Jackson 1984). These are materials that have been mixed from various components by geomorphic processes such as weathering (residual soil), mass wasting (colluvium), glacier transport (till or ice contact deposits), explosive volcanism (granular pyroclastic deposits), or human activity (e.g., fill or mine spoil). Texturally,

Table 2 Landslide velocity scale (WP/WLI 1995 and Cruden and Varnes 1996)

Velocity class	Description	Velocity (mm/s)	Typical velocity	Response[a]
7	Extremely rapid	5×10^{3}	5 m/s	Nil
6	Very rapid	5×10^{1}	3 m/min	Nil
5	Rapid	5×10^{-1}	1.8 m/h	Evacuation
4	Moderate	5×10^{-3}	13 m/month	Evacuation
3	Slow	5×10^{-5}	1.6 m/year	Maintenance
2	Very slow	5×10^{-7}	16 mm/year	Maintenance
1	Extremely Slow			Nil

[a] Based on Hungr (1981)

Table 3 Landslide-forming material types

Material name	Character descriptors (if important)	Simplified field description for the purposes of classification	Corresponding unified soil classes	Laboratory indices (if available)
Rock	Strong	Strong—broken with a hammer		UCS>25 MPa
	Weak	Weak—peeled with a knife		2<UCS<25 MPa
Clay	Stiff	Plastic, can be molded into standard thread when moist, has dry strength	GC, SC, CL, MH, CH, OL, and OH	I_p> 0.05
	Soft			
	Sensitive			
Mud	Liquid	Plastic, unsorted remolded, and close to Liquid Limit	CL, CH, and CM	I_p>0.05 and I_l>0.5
Silt, sand, gravel, and boulders	Dry	Nonplastic (or very low plasticity), granular, sorted. Silt particles cannot be seen by eye	ML	I_p<0.05
	Saturated		SW, SP, and SM	
	Partly saturated		GW, GP, and GM	
Debris	Dry	Low plasticity, unsorted and mixed	SW-GW	I_p<0.05
	Saturated		SM-GM	
	Partly saturated		CL, CH, and CM	
Peat		Organic		
Ice		Glacier		

debris is a mixture of sand, gravel, cobbles and boulders, often with varying proportions of silt and clay. Mud is a similar unsorted material, but with a sufficient silt and clay content to produce plasticity (cohesiveness) and with high moisture content. Both may contain a proportion of organic matter (e.g., Swanston 1974) and may be gap-graded ("diamictons"). Many descriptions found in the literature make reference to coarse clasts and matrix, although no formal separation between these two phases has yet been established. Most often, matrix is considered to be material of sand size or finer, although gravel sizes are sometimes included (Hungr et al. 2001).

Apart from the textural definition in the preceding paragraph, the word "debris" is also traditionally used to describe any material displaced by a mass movement. This wider meaning of the term is not a part of the proposed classification.

An important aspect of debris or mud involved in landslides is that their water content may have been modified by mixing with surface water during motion and could thus be significantly different from the water content of the source material. It may also vary during motion. Spatial gradational sorting of such materials due to the development of inverse grading or coarse surge fronts is common and may have an important bearing on the flow behavior (e.g., Pierson 1986).

Varnes' (1978) criterion that debris is all material containing more than 20 % sizes coarser than sand is probably too restrictive, while at the same time it could apply to plastic and non-plastic materials of widely different characteristics (see Hungr et al. 2001).

Hungr et al. (2001) proposed that the term "mud" be used for remoulded mixed clayey soils whose matrix (sand and finer) is significantly plastic (Plasticity Index>5 %) and whose Liquidity Index during motion is greater than 0.5 (i.e., they are in or close to a liquid state). To convert insensitive stiff or dry cohesive soil at a landslide source into mud, rapid mixing with surface water and increase in porosity is required. Such a mechanism is not often available in nature and this limits the origins of mud to certain specific geological scenarios. For example, many of the mud flows described by Bull (1964) from the desert regions of southwestern USA, contain smectitic clays likely to exhibit dispersive behavior. The word "mud" should not be used to describe remoulded or liquefied clays or silts, which are well sorted and liquefy at their original water content, often without significant mixing with water or other materials.

The word "earth" does not have established status in either geological or geotechnical material description schemes and its use invites confusion with the conventional meaning of earth as construction material or agricultural soil (Bates and Jackson 1984). However, it is required as part of the established term "earth flow." In this context, it means a cohesive, plastic, clayey soil, often mixed and remoulded, whose Liquidity Index is below 0.5. Many earthflows contain fragments of material in different stages of remolding and may carry granular clasts (Keefer and Johnson 1983). To avoid a landslide name that is associated with an unsuitable material term, the term "earthflow" is preferred in this paper (see also Bates and Jackson 1984).

"Ice" was considered a landslide-forming material by Sharpe (1938) and should be re-introduced in the present classification. Many important and destructive mass movements on mountain slopes contain varying proportions of glacial ice and some are dominated by it.

Snow can be an important catalyst for soil saturation and an important cause of rapid motion in some landslides. However, it is not included here among primary landslide-forming materials, to maintain separation from the field of snow science.

Geological material types classified by origin

During preliminary studies, geomorphological analysis often precedes geotechnical testing. Genetic terms can thus add valuable information and can easily be appended to the landslide name, if relevant (Table 4).

Table 4 Supplementary material terms based on geomorphological analysis

Rock	Intrusive, volcanic, metamorphic, strong sedimentary, (carbonatic or arenaceous) and weak sedimentary (argillaceous)
Soil	Residual, colluvial, alluvial, lacustrine, marine, aeolian, glacial, volcanic, organic, random anthropogenic fills, engineered anthropogenic fills, mine tailings, and sanitary waste

However, it is not recommended to replace the material names of the first column of Table 3 by geological terms, because there is often not sufficient equivalency between them. For example, an alluvial deposit may contain clay, silt, sand, or coarser materials. The goal is to stress the component that is the most important in determining the mechanical behavior of the landslide during and post-failure.

Certain geological materials lie on the boundary between soil and rock. Particularly important are saprolites, which combine soil-like physical properties with relict rock mass structure of joints, weathered shear surfaces and similar. Many good reviews of landslides in residual soils exist (e.g., Lacerda 2007). In a universal classification, the authors consider that a landslide in saprolite can be sufficiently well described using one or two of the standard terms selected with the judgment of the user, supplemented by the term "residual soil." A similar approach can be used for highly weathered or mechanically disturbed rock masses.

The following are some examples of landslide names, with assumed supplementary terms:

- Debris slide (residual soil)
- Rock compound slide (weak sedimentary rock)
- Silt flowslide (aeolian silt)
- Clay rotational slide (soft lacustrine clay)
- Clay flowslide (sensitive marine clay)
- Earthflow
- Sand flow (dry fluvial sand)
- Debris flow
- Mud flow
- Debris avalanche (volcanoclastic debris)
- Rock avalanche (strong igneous rock)

Failure distribution and style

Landslide failure mechanisms can be complicated by interaction of adjacent sliding bodies in a variety of styles and distributions. Cruden and Varnes (1996) summarized a number of related descriptors, such as advancing, enlarging, retrogressive, multiple, or successive. Illustrations of some of these terms are shown by Hutchinson (1988). Such terms are a useful supplement to landslide type names.

The term "progressive" is often misused in landslide literature and should be reserved for the specific phenomenon of progressive failure, used in stability or deformation analyses (e.g., Morgenstern 1992; Leroueil et al. 2012).

Another useful group of supplementary terms proposed by Cruden and Varnes (1996) relates to the post-failure activity of the landslide, including re-activated, dormant, and relict.

Definitions of landslide types

General

The following definitions of landslide types are based on Varnes (1978), Hutchinson (1988), Hungr et al. (2001), and other publications. The definitions are supplemented by examples, references, and discussion.

The soil type names presented in italics and separated by a slash symbol are placeholders and only one or two should be used in forming the landslide name.

Falls and topples

1. *Rock/ice* fall: Detachment, fall, rolling, and bouncing of rock or ice fragments. May occur singly or in clusters, but there is little dynamic interaction between the most mobile moving fragments, which interact mainly with the substrate (path). Fragment deformation is unimportant, although fragments can break during impacts. Usually of limited volume.

Detachment of rock fragments from cliffs occurs by a range of mechanisms described under the sliding and toppling categories, occurring at limited scale. Tensile, bending, and buckling failures also play a role. The important distinction of a "fragmental" rock fall (Evans and Hungr 1993) is that individual fragments move as independent rigid bodies interacting with the substrate by means of episodic impacts (Fig. 2). By contrast, rock avalanches (type 18) move in a flow-like manner as masses of fragments. Fragmental rock fall movement can be simulated by numerical models based on rigid body ballistics (e.g., Turner and Schuster 2013).

There is a transition between rock avalanching and rock fall and some events exhibit the character of both. For example, the rock material released by a medium-sized limestone rock wedge slide in Fig. 3 deposited partly as a dry frictional flow of a mass of rock fragments, covering the surface of a talus cone (Bourrier et al. 2012). However, several large fragments decoupled from the depositing mass and bounced and rolled for 300 additional meters in the manner of a fragmental rock fall. The fragment motion being the most dangerous, it is appropriate to call the entire event rock fall.

Given the occurrence of both modes of motion, flow, and rolling/bouncing, a simple definition of fragmental rock fall is

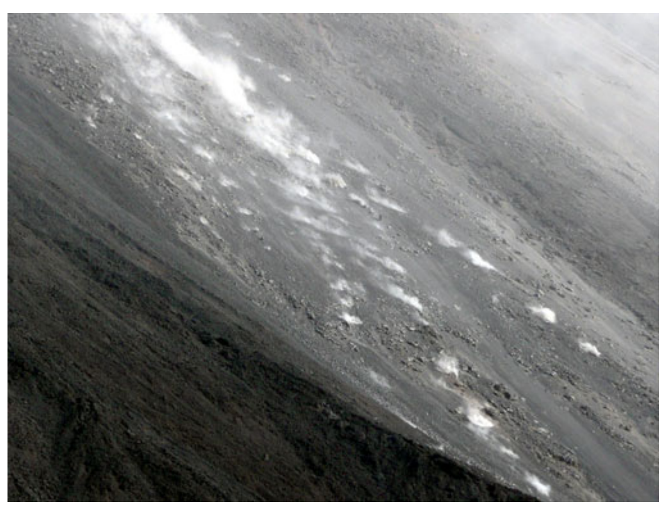

Fig. 2 Rock fall: rock fragments bouncing and rolling over the surface of a talus cone on Mt. Stromboli, Italy (Photo by O. Hungr)

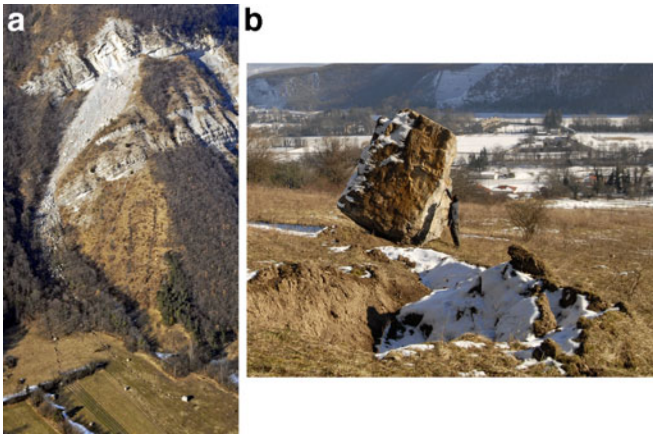

Fig. 3 A rock fall of 1,000 m^3 at St. Paul de Varces, Isère, France. A large part of the unstable mass deposited as a frictional flow on the surface of the talus. Several large boulders rolled approximately 300 m beyond the limits of the granular deposit as shown in the inset (Courtesy of Sébastien Gominet, Institut des Risques Majeurs, Grenoble)

problematic. Some authors attempted a definition based on a maximum volume (e.g., Whalley 1984 proposed 10,000 m^3), but it is difficult to establish a fixed boundary. The suggested recognition criterion is that the most important displacements in terms of distance and hazard intensity should occur in the individual fragment motion mode and can be analysed as such.

Fragmental fall of ice blocks from glacier fronts or icefalls is a common phenomenon associated with alpine glaciers. It is mechanically identical to rock fall, except for the relative weakness of ice blocks. Even small ice falls commonly disintegrate and become granular ice avalanches.

2. *Boulder/debris/silt* fall: Detachment, fall, rolling and bouncing of soil fragments such as large clasts in soil deposits, or blocks of cohesive (cemented or unsaturated) soil. The mechanism of propagation is similar to rock fall, although impacts may be strongly reduced by the weakness of the moving particles.

Soil-derived falls are an important source of hazards on artificial cut slopes along highway cuts and other excavations, or on naturally eroded scarps. The source may be a large clast detached from the soil deposits, or a coherent block of soil. A special category of boulder falls involves the detachment of core stones separated from saprolitic slopes or tors in deeply weathered terrain (ERM-Hong Kong 1998). Core stones, being large and rounded by weathering, can form highly mobile projectiles. They can be released from saprolite surfaces, as the finer material (grus) is eroded around them.

3. Rock block topple: Forward rotation and overturning of rock columns or plates (one or many), separated by steeply dipping joints. The rock is relatively massive and rotation occurs on well-defined basal discontinuities. Movement may begin slowly, but the last stage of failure can be extremely rapid. Occurs at all scales.

The distinction between "block toppling" and "flexural toppling" was introduced by Goodman and Bray (1976). Probably the most important difference between the two types is that block toppling, relying on the rotational stability of thick blocks supported primarily by compressive stress on their bases, is a brittle process: the greater the toppling inclination, the lower the stability, until the point when a sudden acceleration takes place.

Block rotation is often initiated by water pressure in tension cracks, yielding of a weak foundation, or by earthquake acceleration. A classic case of a single block topple was described by Schumm and Chorley (1964). A column of sandstone at the edge of the Chaco Canyon in New Mexico gradually inclined due to slow deformation of a shale foundation. After a decade of extremely slow movements, the 20-m high block suddenly accelerated and overturned within a few seconds, destroying an archeological site. Rock crushing can sometimes be observed at the base of large toppling blocks, as seen on a 2005 video of the failure of the Zenziyan cliff near Chongqing, China (Prof. Y.P. Yin, China Geological Survey, Beijing, personal communication).

The same process can also affect series of blocks separated by steep discontinuities, combined with shallowly dipping joints (Fig. 4). There is, of course, a certain amount of friction on the steep surfaces between adjacent blocks. However, in the case of block toppling, these frictional forces are less important than the stabilizing stresses acting on the bases of the blocks.

Multiple block rotation can accompany sliding in large slopes of strong rock with several joint sets. Brittle failure can occur under certain conditions (Nichol and Hungr 2002). Cruden and Hu (1992) describe a case of toppling of massive calcareous blocks with a bedding dip of 65–70° into the slope and longitudinal joints

Fig. 4 Block topple in limestone, Czech Republic (Photo by O. Hungr)

inclined at about 25° downslope. A mass of 6 million m^3 separated from the slope catastrophically, either as an unstable culmination of the block toppling process, or as a planar slide exploiting the relatively steep basal surface. Cruden and Hu (1992) analyzed the process using a multiple block toppling model proposed by Goodman and Bray (1976). The initiation mechanism of the Mystery Creek rock avalanche, involving 40 million m^3 of intrusive rocks of the British Columbia Coast Ranges, was similarly analyzed by Nichol and Hungr (2002). Such landslides may be called block topples in the rather rare cases where toppling is the dominant mechanism, but may also be termed rock slides (types 7, 9, or 10).

4. Rock flexural topple: Bending and forward rotation of a rock mass characterized by very closely spaced, steeply dipping joints or schistose partings, striking perpendicular to the fall line of the slope. The rock is relatively weak and fissile. There are no well-defined basal joints, so that rotation of the strata must be facilitated by bending. The movement is generally slow and tends to self-stabilize. However, secondary rotational sliding may develop in the hinge zone of the topple. Occurs at large scale.

Flexural toppling is a fundamentally different process. The major principal stress near the face surface of large slopes is oriented parallel with the slope face. A kinematic criterion devised by Goodman and Bray (1976) for closely jointed rock slopes shows that, if both the dip of the joints and the slope inclination are steep enough, reverse slip can occur along the controlling joints. The thin layers of weak rock bend in the downslope direction. Characteristic reverse scarps form on the slope surface, as can be seen by the vegetation-enhanced horizontal lineaments crossing the upper part of the slope in Fig. 5a. As the rock strata rotate forward, shear stresses on the column or plate sides resist movement. The magnitude of these stresses increases with forward rotation and the mechanism is, therefore, self-stabilizing. Thus, in a marked contrast to block toppling, flexural toppling tends to be a slow, ductile process (Nichol and Hungr 2002).

Flexural topples can occur both in anaclinal and cataclinal slopes (Cruden 1989). If flexural deformation occurs at depth, a related mechanism termed kink band slumping results (Kieffer 2003). This is transitional to slope deformation movements (Type 28).

There are many examples of slow toppling movements of large mountain slopes. The zone of maximum bending curvature of the strata ("the hinge zone") can develop a shear band and the landslide may thus evolve into a rotational slide (type 6), as occurred in the central portion of the La Clapière slope shown in Fig. 5a, b (Follacci 1987). Partial detachments of this type may reach catastrophic movement rates (Chigira and Kiho 1994).

5. *Gravel/sand/silt* block topple: Block toppling of columns of cohesive (cemented) soil, separated by vertical joints.

The mechanism of block toppling in soil is equal to block toppling in rock, although the low strength of weakly cemented or of partially saturated soil columns promotes failure by basal crushing, without the need for horizontal discontinuities. Hungr et al. (2001) describe dry silt flows caused by toppling of columns in jointed, cohesive glacio-lacustrine silt of the British Columbia Interior, Canada, which have produced dry flows sufficiently mobile to destroy houses and cause fatalities.

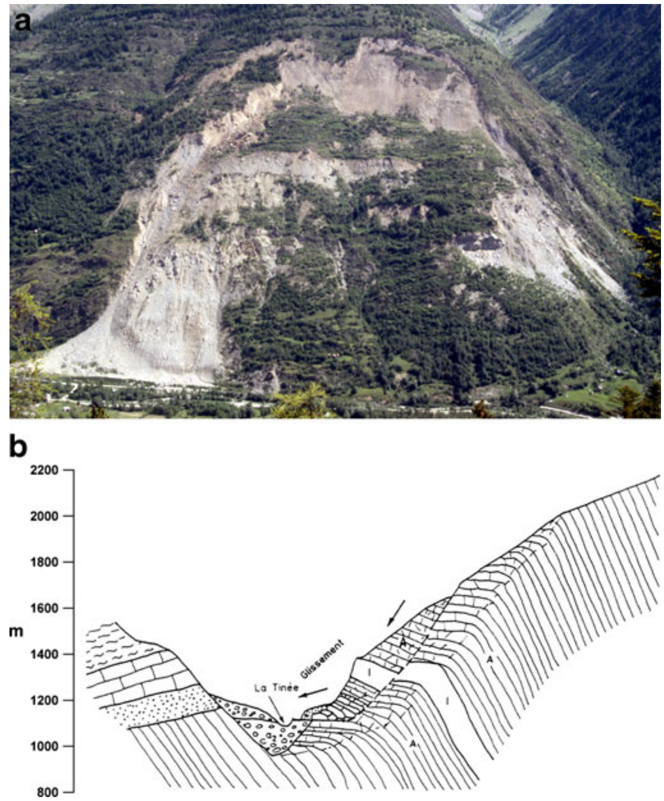

Fig. 5 **a** Large flexural topple, in the process of converting into a rotational slide. La Clapiére, France (photo, O. Hungr). **b** Schematic cross-section of the Clapiére flexural topple by Follacci (1987). A micaceous gneiss, I quartzitic gneiss

Slides in rock

6. Rock rotational slide ("rock slump"): Sliding of a mass of weak rock on a cylindrical or other rotational rupture surface which is not structurally controlled. The morphology is characterized by a prominent main scarp, a characteristic back-tilted bench at the head and limited internal deformation. Usually slow to moderately slow.

Rotational slides can occur only in very weak rock masses, often under the surcharge of a stronger cap rock (Fig. 6). Most rotational slides in rock tend to move at slow or moderate velocities, partly because the rotational mechanism is self-stabilizing as the gravitational driving forces diminish with increasing displacement. More importantly, weak rock mass under shear stress tends to fail in a ductile manner (Hungr and Evans 2004a). The reasons for the ductile behavior often displayed by landslides in weak rocks are complex and poorly understood. It is possibly a consequence of pre-failure progressive deformations which destroy the cohesion before general failure is attained.

However, there are some exceptions. Rotational sliding of weak rock surcharged by a thick cap of strong, brittle rock can

Fig. 6 A rotational slide involving Cretaceous shale, overlain by sandstone. Liard Plateau, Canada (Photo by O. Hungr)

sometimes induce extremely rapid rock avalanches, as parts of the cap rock topple or slide along discontinuities and impact clayey debris accumulated on the lower slope surface. A spectacular example is the 1915 Great Fall of Folkestone Warren where three blocks of chalk, destabilized by ductile rotational sliding of the underlying clay shale, fragmented and swept over the disturbed lower slopes at extreme speed (Hutchinson et al. 1980). A similar landslide at Roccamontepiano in the central Apennines, was the fourth deadliest landslide in European history (D'Alessandro et al. 2002). The site is on the edge of a 300-m high table mountain built of overconsolidated Tertiary marine clay, capped by a 40-m thick layer of travertine. After visible deformation and smaller precursory failures, a portion of the travertine cliff collapsed on 24 June 1765 and a rock avalanche swept for 2 km over the slope, destroying a village and causing 700 fatalities. It is of interest to note that brittle, but porous cap rock was involved in each of these cases, possibly promoting the mobility of the resulting rock avalanches (cf. Hutchinson 2002).

A very unusual case of rapid rotational sliding was the large landslide of February, 2010 at Maierato in Calabria, Italy (Guerricchio et al. 2012). This is a slope in Miocene clays and calcareous beds, which had failed extensively by deep-seated rotational sliding during an earthquake in 1783. Heavy rains in early 2010 re-activated a part of the unstable masses, involving approximately 10 million m^3, along a roughly rotational surface. A video available on the web shows surface movements in the range of several meters per second in the center of the displaced mass. The high velocity and flow-like character of the landslide suggests that increase of pore-pressure took place, so that the climax of the movement seen on the video could also be termed a flowslide (type 20 below).

7. Rock planar slide ("block slide"): Sliding of a mass of rock on a planar rupture surface. The surface may be stepped forward. Little or no internal deformation. The slide head may be separating from stable rock along a deep, vertical tension crack. Usually extremely rapid.

Some of the largest and most damaging landslides on Earth are translational landslides, such as the prehistoric Seimareh slide in the Zagros Mountains of Iran (Roberts and Evans 2013), or the Flims rock slide in the Alps (Heim 1932). However, planar rock slides occur at all scales in layered, folded sedimentary rocks, metamorphic rocks which fail along schistosity or fault planes and in intrusive rocks with stress relief joints (exfoliation).

The planar sliding mechanism is not self-stabilizing and the slides tend to be extremely rapid, except in the case of very weak rocks and failures on very flat-dipping discontinuity planes. Sometimes, a minor dip of the strata can have spectacular results. The 1248 rock avalanche at Mt. Granier, in the Savoy Alps, was the deadliest landslide in European history, destroying a regional town with some 5,000 inhabitants. Figure 7, based on an interpretation by Cruden and Antoine (1984), shows that the landslide occurred in a sedimentary sequence with a downslope dip of some 12° to 17°. The landslide block, over 200 million m^3 in volume (Goguel and Pachoud 1972), detached from a vertical side scarp, probably rotated around a vertical axis, disintegrated and transformed into a rock avalanche travelling for 7 km. This brittle behavior contrasts strikingly with the ductile failure of the Massif de Platé rotational slide described by Goguel and Pachoud (1981), despite the fact that both events occurred in virtually identical geological settings and were of comparable volume. Only the moderate downslope dip

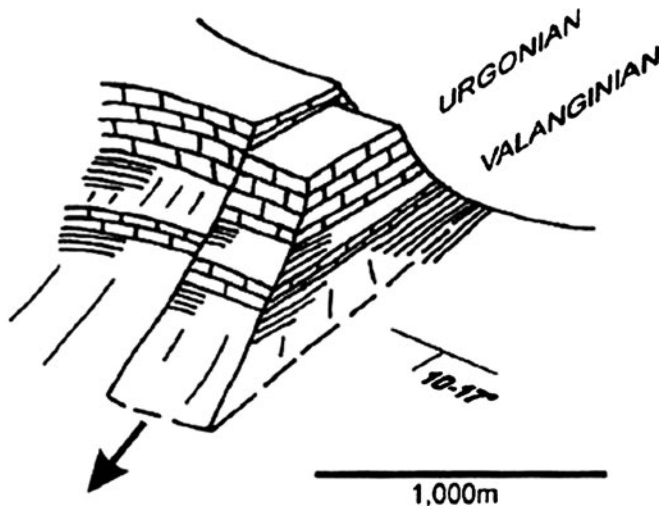

Fig. 7 Mont Granier translational rock slide. Schema of the failure mechanism, based on an interpretation by Cruden and Antoine (1984). The vertical surface of separation behind the moving block is a tension feature

facilitated brittle, translational detachment in the case of Mt. Granier (Hungr and Evans 2004a).

Planar slides usually involve dip slopes that have been undecut by erosion or excavation. In some cases, the undercutting is not complete and a "toe breakout" mechanism involving failure of intact rock mass must develop. The abovementioned Seimareh landslide provides a spectacular example (Roberts and Evans 2013).

It is important to differentiate between translational slides and compound slides with a bi-linear rupture surface (e.g., Hutchinson 1988). The latter consist of an "active" proximal block, driving a relatively stable "passive" or stabilizing block (see below). By contrast, in true translational slides, the entire sliding body is in an active state and its proximal margin (if any) fails in tension. The tall vertical head scarp of Mt. Granier is therefore not the base of an active sliding block, but simply an open tension surface.

8. Rock wedge slide: Sliding of a mass of rock on a rupture surface formed of two planes with a downslope-oriented intersection. No internal deformation. Usually extremely rapid.

Wedge slides are translational slides exploiting favourably oriented intersecting discontinuities (Fig. 8). Mechanically, wedge slides are analogous to planar sliding, except that the stabilizing forces are increased by a wedge factor, being a function of the attitude of the controlling planes, as well as the strength properties of the discontinuities and pore-pressures (Hoek and Bray 1981). They occur at a range of scales, although most are small.

9. Rock compound slide: Sliding of a mass of rock on a rupture surface consisting of several planes, or a surface of uneven curvature, so that motion is kinematically possible only if accompanied by significant internal distortion of the moving mass. Horst-and-graben features at the head and many secondary shear surfaces are typical. Slow or rapid.

Fig. 8 Scars of wedge failures in limestone, Canmore, Alberta, Canada. The cliff is approximately 50 m high (Photo by O. Hungr)

The most common type of a compound rock slide has a rupture surface following a horizontal, or gently inclined plane of weakness such as a bedding plane or a weak layer in the stratigraphy, daylighting at the toe (Hutchinson 1988). A steep main scarp cutting through the rock mass forms the proximal part of the rupture surface, to daylight at the crown. The shape of the rupture surface in profile may be bi-linear or curved (listric), but noncircular. An example of a compound slide controlled by a weak bedding plane in Lower Cretaceous shales is shown in Fig. 9a. Note the horst and graben structure at the head of the slide, indicating a nearly horizontal displacement of the horst block along the basal surface. A back analysis of the landslide showed that the effective friction angle on the weak surface, likely formed by a pre-sheared bentonite seam, must be only 8°. As shown in Fig. 9b, small regional dip of the bedding appears to have caused the marked asymmetry of the valley (Gerath and Hungr 1993).

Compound geometry may in some cases be formed by the curvature of tectonic folds. A key example is the 1963 Vaiont Slide, where an extensive rupture surface, seated on clay-coated bedding planes in limestone is shaped along a curving, but noncylindrical, plunging syncline (Hendron and Patton 1985). Sliding movement along this surface requires internal deformation of the landslide body, engaging the high strength and brittleness of the limestone rock mass (Mencl 1966; Hutchinson 1988). In slope stability analysis of such cases, the mobilized internal strength of the sliding body must be taken in consideration.

10. Rock irregular slide ("rock collapse"): Sliding of a rock mass on an irregular rupture surface consisting of a number of randomly oriented joints, separated by segments of intact rock ("rock bridges"). Occurs in strong rocks with non-

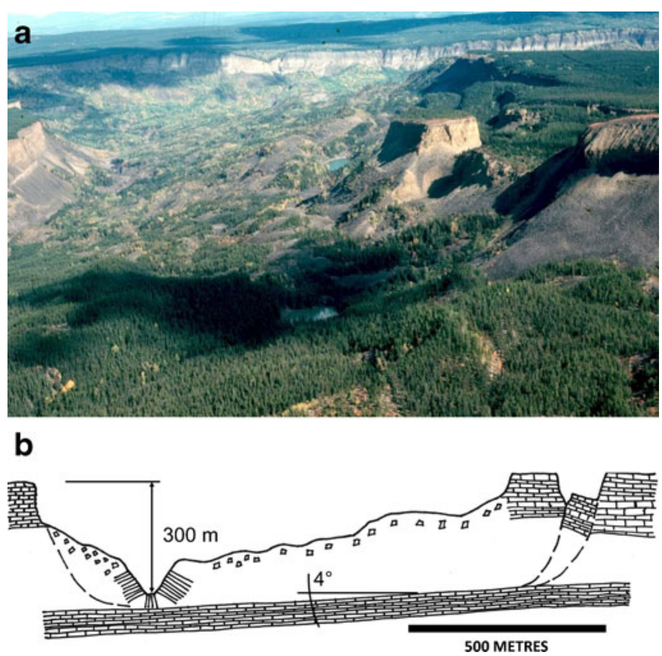

Fig. 9 **a** A compound slide in Cretaceous shale, Liard Plateau, British Columbia. **b** Schematic cross-section (Gerath and Hungr 1993). The valley is an ancient meltwater channel, occupied by a grossly underfit stream

systematic structure. Failure mechanism is complex and often difficult to describe. May include elements of toppling. Often very sudden and extremely rapid.

Rock slides originating on steep slopes in strong rock often fail by a complex mechanism, exploiting a number of discontinuities separated by segments of intact rock mass. The existence of discontinuities is essential, because strong rock at shallow depths is rarely stressed close to failure. However, the rock structure is not systematic, so it is not possible to place the failure mechanism into any of the three preceding categories (Fig. 10). The rupture surface forms by connecting many randomly oriented and non-persistent discontinuities, separated by segments of intact rock ("rock bridges"). The overall shape of the rupture surface is irregular and kinematically complex to a varying degree. Parts of the sliding mass may simultaneously be toppling.

The stability of the slope relies primarily on the extent and strength of the rock bridges, which fail by means of propagation of shear or tensile cracks. As the geometry and properties of rock bridges are usually unknown, a meaningful slope stability analysis cannot be carried out and the only way to manage a developing landslide of this type is by the use of the observational method. An excellent example is the 1991 Randa rock slide in the Zermatt Valley, Switzerland, where very detailed site investigation and geophysical observations permitted a prediction of failure and outlined the eventual complex failure mechanism (Eberhardt 2008). Another extensively studied example is the Séchillienne landslide near Grenoble, France which has been moving for several decades and produced numerous minor detachments, but has so far avoided forming a clear and identifiable pattern of overall failure (Antoine et al. 1987). Many potentially catastrophic rock slides in strong rock belong to this challenging category.

Slides in soil

11. *Clay/silt* rotational slide ("soil slump"): Sliding of a mass of (homogeneous and usually cohesive) soil on a rotational rupture surface. Little internal deformation. Prominent main scarp and back-tilted landslide head. Normally slow to rapid, but may be extremely rapid in sensitive or collapsive soils.

Purely rotational slumps with a cylindrical or ellipsoidal shape of the sliding surface in cohesive soils are probably more common in soil mechanics textbooks than in nature. Under natural conditions, the shape of the rupture surface usually departs to a certain degree from constant curvature, tending towards compound sliding (type 14 below).

Fig. 10 Rock collapse, Preonzo, Switzerland (Courtesy of S. Löw, ETH, Zurich)

Deep-seated rotational motion is favoured by undrained failures and is most common in saturated soils of low permeability (clays or silts). Under special circumstances, undrained failure can occur in granular soils, particularly if rapid motion is triggered by liquefaction. In such cases, the circular sliding failure mode is short-lived and serves as an initiating mechanism of flowslides (type 20).

Rotational slides occur in homogeneous massive clay deposits excavated by stream erosion or artificial earth works and in man-made fill slopes. Morphologically, a soil slump is characterized by a prominent main scarp ("head scarp") and a back-tilted bench forming the head of the slide. The body of the slide usually shows some, but limited, internal deformation. The movement is commonly slow or moderate in velocity, unless the clay is sensitive.

Initial rotational slides in very sensitive clays are extremely rapid and produce remolding accompanied by an extremely high degree of strength loss. As the liquid remolded clay flows away from the steep initial main scarp and removes support from it, a retrogressive slide results. This process may repeat itself many times in a multiple-retrogressive fashion, forming sensitive clay flowslides, as described under type 21 below. In these cases, rotational sliding is merely an initial stage of a more important slope movement and a mechanism of retrogression.

12. *Clay/silt* planar slide: Sliding of a block of cohesive soil on an inclined planar rupture surface, formed by a weak layer (often pre-sheared). The head of the slide mass separates from stable soil along a deep tension crack (no active wedge). May be slow or rapid.

Shearing failure in cohesive materials prefers curved rotational or compound sliding surfaces. If a planar slide occurs, it is likely controlled by a weak layer or a discontinuity, inclined at an angle exceeding the friction angle (with an allowance for pore-pressure and earthquake body forces). Some of the famous slides in clay shale of the Gaillard Cut of the Panama Canal were of this type, being controlled by pre-sheared bentonitic seams (Lutton et al. 1978). Figure 11 shows a spectacular case of a planar slide in a Tertiary clay characteristic of the Piedmont region of northern Italy (Forlati et al. 1998). As in many other slides in weak shales or overconsolidated clays, the failure surface follows a bedding discontinuity, pre-sheared to residual friction. The pre-shearing could be tectonic in origin, although progressive failure may also play a role (Morgenstern 1992). Large planar slides mobilized on tectonically pre-sheared fault surfaces oblique to bedding have also been described in the tectonized "varicoloured clays" of the Italian Apennine foothills (Bozzano et al. 2008).

Recognizing the important role of pre-shearing in planar and compound soil slides, the classification system of Sassa (1999), included a special type for landslides sliding on surfaces at residual friction.

A special type of extremely rapid planar slide involving a block of insensitive clay overlying a thin layer of very sensitive clay was described by Hutchinson (1961) from Furre, Norway. Hutchinson proposed the term "flake slide" in quick clay. In the classification of very sensitive clay landslides proposed more recently by Locat

Fig. 11 A translational slide in tectonized Tertiary clay shale, Murazzano, Langhe district, northern Italy (Courtesy of Servizio Geologico della Regione Piemonte and C. Scavia, Turin Polytechnic)

et al. (2011), the same phenomenon is referred to as a "translational progressive landslide." A remolding process driven by progressive failure is postulated as the controlling mechanism. As is the case for other landslides involving very sensitive clay, the failure is extremely rapid.

13. *Gravel/sand/debris* slide: Sliding of a mass of granular material on a shallow, planar surface parallel with the ground. Usually, the sliding mass is a veneer of colluvium, weathered soil, or pyroclastic deposits sliding over a stronger substrate. Many debris slides become flow-like after moving a short distance and transform into extremely rapid debris avalanches.

Dry, homogeneous granular soil is close to an ideal frictional medium and tends to fail as a thin layer of instability, just beneath the ground surface standing at the angle of repose. Such planar sliding is the initial phase of a dry granular flow, such as is often observed on stockpiles of sand, or on the lee slopes of sand dunes (type 19).

Thin veneers or blankets of loose, poorly sorted soil covering steep slopes formed of a stronger substrate are very common in all mountainous and hilly regions of the world. The most typical are colluvial veneers of soil disturbed and transported by soil creep and vegetation activity, overlying denser and stronger soil deposits, or bedrock. Examples of these are shown below as initial failures for debris avalanches (type 25).

Some characteristics of debris slides initial to debris avalanche clusters are fairly consistent among different regions. For example, natural debris slides in Venezuela (Larsen and Wieczorek 2006), British Columbia (Jakob 2000), and Hong Kong (Dai and Lee 2003), are 0.5 to 2 m thick and initiate primarily on angles of 30–60°, with less than 8 % initiating between 20° and 30°. Slope angles greater than 30° reflect relatively high friction angle of colluvial veneers, augmented by true cohesion (cementing), apparent cohesion (due to incomplete saturation) and binding action of root systems, all of which occur in near-surface soils. Sliding is rare on slopes steeper than about 60°, as such slopes do not tend to support soil veneers. Some debris slides exploit smooth interfaces between strong bedrock and the colluvial veneer (e.g., Lacerda 2007) and others occur in residual weathered horizons or paleosols (Guadagno et al. 2005).

In humid tropical or temperate regions, organic soil veneers cover steep bedrock and are prone to detachment and sliding. Examples from Hawaii have been shown by Cannon (1993) and others. In regions close to centers of explosive volcanism, such as the Campania Region of Italy, or the area north-east of Mt. St. Helens, USA, an unstable surficial veneer is formed of pyroclastic deposits over steep bedrock slopes (e.g., Guadagno et al. 2005; Picarelli et al. 2008).

Such veneers are highly susceptible to landslides, for several reasons: (1) The veneers are much weaker than the underlying material and are able to persist on steep slopes only by virtue of cementing, negative pore-pressures due to incomplete saturation, or vegetation root reinforcement. When any of these factors are decreased, instability occurs. (2) In many cases, the interface between the veneer and the substrate is smooth and therefore weaker than the soil itself. (3) The contrasting permeability of the veneer and substrate may promote rapidly recharging perched water tables and slope-parallel flow, or destabilizing upward seepage. Workers dealing with permafrost slides refer to shallow sliding of the active layer, overlying frozen ground as "skin flows" (e.g., McRoberts and Morgenstern 1974).

Shallow planar slides are most frequently triggered by extreme rainfall. For example, the deadly 1999 debris avalanches in the Vargas Province of Venezuela took place during rainfalls exceeding 900 mm over a 3-day period and over 400 mm/24 h (Larsen and Wieczorek 2006). It is common to observe spatial correlation between storm rainfall intensity contours and landslide density (e.g., Crozier 2005; Guthrie and Evans 2004; Coelho Netto et al. 2011). Removal of vegetation by logging or fire tends to increase both the density of debris slides and the amount of material moved (Cannon and Gartner 2005; Jakob 2000).

Saprolitic or lateritic residual soil profiles in particular, can be substantially weaker than the underlying parent material. Steep saprolite slopes in Brazil, for example, have special characteristics that promote planar failure (Lacerda 2007). These characteristics include relict joint planes (including slope-parallel exfoliation joints), apparent cohesion due to suction, which can be destroyed by the downward advance of a wetting front, or the formation of artesian pressures in saprolite, topped by an impervious lateritic horizon. Saprolites are also often lightly cemented by oxides and the resulting true cohesion may be destroyed by repetitive wetting and drying.

Theoretical analysis of coupled groundwater seepage and slope stability predicts that the most likely sites for debris slides should be situated in zones where seepage accumulates, such as depressions and floors of gullies (e.g., Montgomery and Dietrich 1994). However, the siting of debris slides depends on many factors and varies with region. For example, debris avalanches in the humid coastal ranges of British Columbia, Canada, most often initiate as artificial fill failures along forestry roads (e.g., O'Loughlin 1972). Many of the debris slides initial to the deadly debris avalanches in 1998 at Sarno, Italy started at slope breaks caused by agricultural roads, or natural cliff bands (Guadagno et al. 2005). Some also initiated in areas where karstic springs open to the slope surface beneath the pyroclastic veneer (Cascini et al. 2008). Saprolite debris slides in eastern Brazil most often initiate at the crests of slopes, probably because they are triggered as tension cracks infill with water during heavy rains (W.A. Lacerda, Federal University of Rio de Janeiro, personal communication, 2011).

As failure of loose veneers on steep slopes involves loss of cohesion and often also partial or full spontaneous liquefaction, the slides usually behave in a brittle manner, accelerate, lose coherence and continue downslope in the form of flow-like debris avalanches (type 25 below). Most debris slides will thus be classified as an initial component of debris avalanches (25) or debris flows (22), to which they serve as a mechanism of initiation. It is rare for a debris slide to fail in a ductile manner and remain near or within the source area.

Much larger translational debris slides occur in accumulations of coarse colluvium, built by rock fall or rock slides on steep slopes. Such accumulations of boulders and sandy matrix can be re-activated by toe erosion, high infiltration and increase of pore-pressure, or earthquake shaking. The soil masses slide forward, sometimes reaching fairly high mobility. These failures can become large and destructive debris avalanches (type 25). The 1.2 million m^3 debris avalanche at Cortenova, Italy, shown in Fig. 12, initiated as a translational slide of previously deposited landslide debris, re-activated following a period of extreme infiltration in 2002 (Crosta et al. 2005). Similar large, destructive movements of bouldery accumulations of colluvial debris were triggered by the 2010 Wenchuan earthquake, with disastrous consequences (Prof Y.P. Yin, pers. comm.).

14. *Clay/silt* compound slide: Sliding of a mass of soil on a rupture surface consisting of several planes, or a surface of uneven curvature, so that motion is kinematically possible only if accompanied by significant internal distortion of the moving mass. Horst-and-graben features at the head and many secondary shear surfaces are observed. The basal segment of the rupture surface often follows a weak horizon in the soil stratigraphy.

Like in rock, compound soil slides form where a weak horizon attracts the major distal part of the rupture surface, while a steep main scarp and a horst-and-graben structure form at the head. Again, sliding along compound surfaces in soil requires strong internal distortion of the sliding body, often resulting in multiple internal shears distributed throughout the slide (Fig. 13).

Compound slides are widespread in glacio-lacustrine deposits of Western Canada, where clay interbeds in silty or sandy strata form the weak layers. West of the Rocky Mountains, where Cretaceous shales underlie glacio-lacustrine deposits, one can often find compound slides of very similar morphology both in bedrock and the Pleistocene soils on multiple levels, often in a successive sequence. In some cases, the weak plane is situated in bedrock, while the main scarp and the horst-and-graben structure form in the overlying soil. In such cases, the user of the classification must decide whether to place the landslide into types 9 or 14.

Spreading

15. Rock slope spread: Near-horizontal stretching (elongation) of a mass of coherent blocks of rock as a result of intensive deformation of an underlying weak material, or by multiple retrogressive sliding controlled by a weak basal surface. Usually with fairly limited total displacement and slow.

Rock slope spreading, involving the displacement and rotation of rigid blocks of stronger rock, because of severe plastic deformation of an underlying layer of weak rock is very common in horizontally bedded, weak sedimentary sequences. A large variety of such landslides has been detailed from the Czech and Slovak Republic in classic books by Zaruba and Mencl (1969) and Nemčok (1982)—Fig. 14, as well as in Southern England (Hutchinson 1991).

Fig. 12 The 2002 debris avalanche at Cortenova, Lombardy, Italy, which initiated as a translational slide of previously disturbed landslide debris derived from metamorphic rocks (Courtesy of G. Crosta, University of Milan, Bicoca)

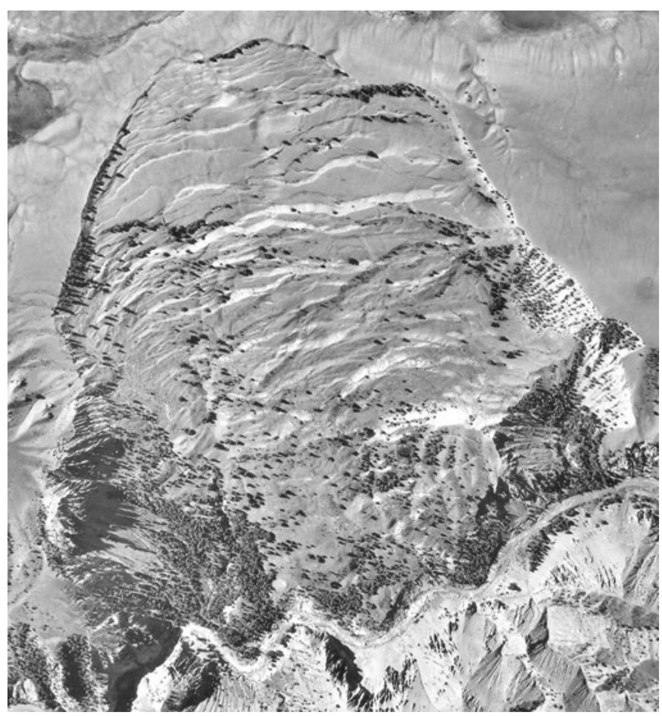

Fig. 13 Vertical aerial photo of a compound slide in glacio-lacustrine deposits, Churn Creek, British Columbia Interior. B.C. Government Airphoto BC7721. The frame is approximately 1 km wide. Note that internal shears form scarps both in normal and anti-slope directions

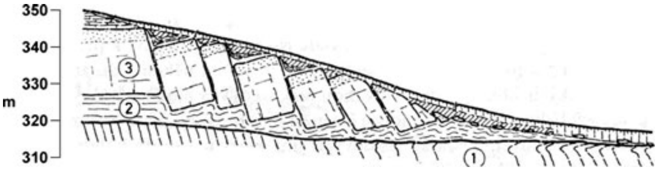

Fig. 14 Slow spreading of blocks of sandstone blocks due to the deep deformation of a weak shale substrate, Prague, Czech Republic: *1*, Phyllite; *2*, deformed shale, and *3*, sandstone. (Zaruba and Mencl 1969)

Many examples have also been observed in Central Italy, where tectonized clay shales deform under the weight of volcanic or sedimentary cap rock (e.g., Canuti et al. 1990; Picarelli and Russo 2004).

This type of spreading is transitional from the rock slope deformation phenomena described below under type 29. The user of the classification must decide between placing a given case into one category or the other. The term "spread" should be applied where a large and well-defined part of the slope has undergone distinct displacements so that a "rupture surface" can be defined, separating the zone which has moved from one that has remained stationary. When the rupture surface consists of a discrete shear plane, or a thin shear band, it is better to speak of a compound slide (type 9). Conversely, "deformation" applies to cases where there is a gradual increase in plastic straining with depth.

Spreading by multiple retrogressive sliding failure is a related process, where a number of instabilities exploit a single weak horizon. Spectacular examples from the valley of the South Saskatchewan River, some 170 m south of Saskatoon, Canada, are shown on Fig. 15a (see also Mollard and Janes 1984, plates 3–40, page 261). A mechanical explanation of the process creating such features was proposed by Haug et al. (1977), as shown schematically on Fig. 15b. This area of Saskatchewan is covered by glacial drift and glacio-lacustrine clay, deposited on a bedrock surface formed by Cretaceous shales, containing bentonite layers. The bentonite horizons are often pre-sheared to residual friction by glacial drag, valley rebound deformation and/or progressive failure. The slope failures probably initiated in late Pleistocene time, when intensive meltwater flows undermined the valley slopes formed of the weak rock and soil. Multiple retrogressive sliding took place, forming the spreading features. Present day retrogression is not very frequent, because the current rivers lack the erosive power of former melt water flows. The movement rates are slow.

16. *Sand/silt* liquefaction spread: Extremely rapid lateral spreading of a series of soil blocks, floating on a layer of saturated (loose) granular soil, liquefied by earthquake shaking or spontaneous liquefaction.

This type of spreading occurs as a result of spontaneous or earthquake liquefaction, where the liquefiable material forms only a small part of the unstable volume. The remainder of the material breaks into more-less intact blocks, which "float" on a mobile layer situated at depth. A classic case occurred during the 1964 Alaska Earthquake in a glacio-marine terrace at Turnagain Heights, Anchorage, Alaska (Fig. 16). The terrace was formed of overconsolidated clay of moderate sensitivity, which broke into large blocks. According to Seed and Wilson (1967), the clay blocks were carried in liquid sand, as a result of liquefaction of loose, saturated sand lenses in the soil profile. Sand volcanoes were observed among the tilted blocks of clay.

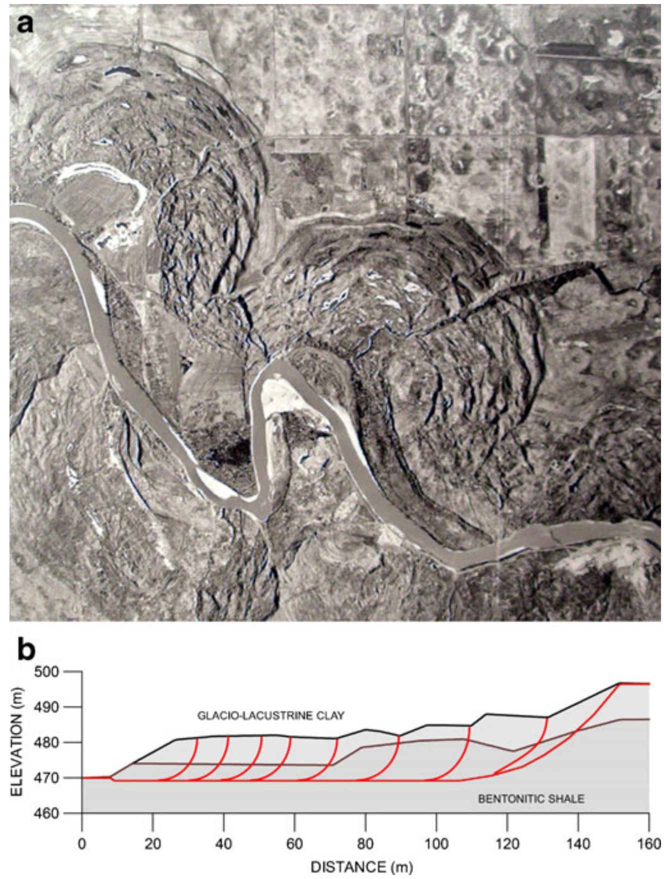

Fig. 15 **a** Lateral spreads caused by multiple-retrogressive compound sliding in glacio-lacustrine clay, overlying Cretaceous shale with pre-sheared bentonite seams in Saskatchewan, Canada. Government of Canada Airphoto 5511–68. The area depicted is approximately 3 km wide. **b** Schematic cross-section through a rock spread similar to that shown in (**a**) (after Haug et al. 1977)

Similarly, the failure of the upstream face of the San Fernando Dam in California during an earthquake in 1971 involved liquefaction of zones of loose sand. The upstream part of the dam cross-section spread laterally from an original width of 90 to 140 m. However, only some 20–30 % of the cross-section material liquefied. The remainder was made up of displaced intact blocks of compacted soil (Seed et al. 1973).

17. Sensitive clay spread: Extremely rapid lateral spreading of a series of coherent clay blocks, floating on a layer of remoulded sensitive clay.

Sensitive clay spreads result from the propagation of a quasi-horizontal shear zone from the toe of the slope (Locat et al. 2011) over which more or less intact soil blocks move laterally towards the valley. The intact blocks may form back-tilted benches (Fig. 17) or series of horsts and grabens (Fig. 18). Most of the displaced material remains in the landslide source area. The scenario is

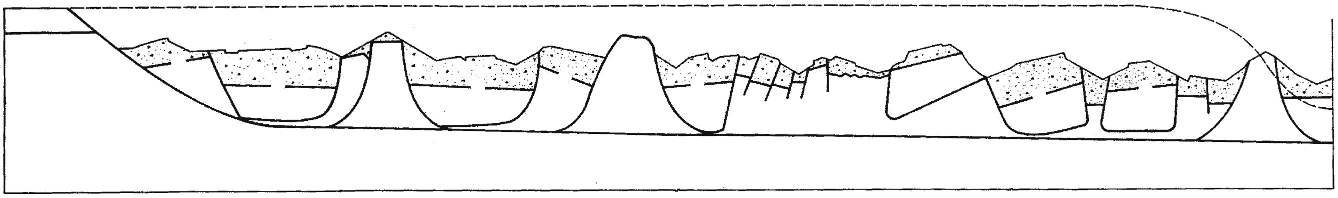

Fig. 16 Interpreted cross-section of the liquefaction spread at Turnagain Heights, Anchorage, Alaska, during the 1864 Alaska Earthquake (Adapted from Seed and Wilson 1967) The dashed line is the original pre-failure surface of the marine terrace. The approximate depth and length of the depicted failure zone are 20 and 120 m respectively

typical of soil profiles involving sensitive marine clay, capped by a stiff desiccated crust.

Although the morphological differences between the two landslide types are fairly obvious, the distinction between a flowslide (type 21) and a spread in sensitive clay (type 17) depends on the relative proportions of liquid and solid material in the landslide, which can often only be estimated by judgment.

Flow-like landslides

18. *Rock/ice* avalanche: Extremely rapid, massive, flow-like motion of fragmented rock from a large rock slide or rock fall.

Large rock slides disintegrate rapidly during motion down mountain slopes and travel as extremely rapid flows of fragmented rock (Fig. 19). Heim (1932) coined the term "sturzstrom" (rockslide stream) to describe these landslides. Beginning with Heim, numerous authors pointed out that large rock avalanches achieve a degree of mobility that far exceeds what would be expected from a frictional flow of dry, angular, broken rock. Furthermore, the mobility increases systematically with volume of the event.

The deposits exhibit rough inverse sorting. The bulk of the rock avalanche mass is dry during motion, because the extensive fragmentation of the rock mass generates very large new pore-space that cannot be filled with water during the short time of motion. However, in many cases observed in the field, the rock avalanche debris travels on a cushion of saturated material entrained from the flow path and liquefied by rapid undrained loading under the weight of the rock debris (Hungr and Evans 2004b). Many alternative explanations of the "excessive mobility" of large rock avalanches have been proposed in the literature, but none has so far gained universal acceptance and a lively discussion continues on this subject.

Some authors have suggested that the term "sturzstrom," implying excessive mobility, should be reserved for events exceeding about 1 million m^3. It is true that most small rock avalanches show moderate mobility that can be readily explained using dynamic models based on frictional mechanics (e.g., Strouth and Eberhardt 2009). However, examples of some very mobile, small rock avalanches have been described in the literature. Their mobility is possibly the consequence of special characteristics of the rock material (e.g. Hutchinson, 2002), or of entrainment of saturated material from the base of the landslide (Hungr and Evans, 2004b). A comprehensive recent review of rock avalanches and their impact has been compiled in Evans et al. 2006.

Hutchinson (2002) proposed a hypothesis that crushing of porous material during failure creates excess pore-pressure within a basal shear band, a theory similar to the "sliding surface liquefaction" concept advanced with laboratory testing support by Sassa (e.g., 2000). The 180 million-m^3 Bairaman rock avalanche on Papua New Guinea is a striking example of such a phenomenon (King et al. 1989). Here, a thick block of porous, karstified Tertiary limestone (Biosparite) was destabilized by a Magnitude 7.1 earthquake on a sliding surface dipping by only a few degrees towards a river gorge. Apparently, the block entirely disintegrated and produced a flowslide containing less than 10 % boulders and moving a distance of over 2 km on an essentially horizontal slope.

Glacier ice is often involved in avalanching of mountain slopes. Ice may form a part, or possibly all of the moving mass or, a rock avalanche can move over the surface of a glacier. The largest recent

Fig. 17 a A lateral spreading failure following rotational sliding in extra-sensitive clay, St. Jude, Quebec, Canada. Photo Ministère des Transports du Québec

Fig. 18 Horst and graben structure at the head of a lateral spread in sensitive clay, St Liguori, Québec, Canada (Courtesy of S. Leroueil)

Fig. 19 "Frank Slide" rock avalanche of 1903, southern Alberta, Canada. The horizontal length of the avalanche path is 3 km, volume 36 million m^3. (Photo by O. Hungr)

glacier failure event, shown in Fig. 20, is the Karmadon-Kolka ice avalanche of 2002 in the Caucasus Mountains. This comprised nearly complete collapse of a valley glacier and avalanching of some 130 million m^3 of fragmented ice over a distance of 19 km, reaching velocities of over 250 km/h (Evans et al. 2009). A small rock avalanche of 3.2 million m^3, running out over the surface of a valley glacier in the Coast Ranges of British Columbia was studied by Delaney and Evans (2013) and is shown in Fig. 21. As noted by the last reference and by many authors previously, avalanches involving glacier ice either as the moving material or as the substrate, demonstrate exceptionally high mobility. The most deadly single landslide accident in history was the tragic, earthquake-triggered Huascaran rock and ice avalanche of 1970, which destroyed a town and caused approximately 15,000 fatalities (Plafker and Ericksen 1978).

19. Dry (or non-liquefied) *sand/silt/gravel/debris* flow: Slow or rapid flow-like movement of loose dry, moist or subaqueous, sorted or unsorted granular material, without excess pore-pressure.

Fig. 20 Path and deposits of the 2002 Kolka Glacier ice avalanche in the Caucasus Mountains (Evans et al. 2009). (Courtesy of O. Tutubalina and S. Chernomorets, Moscow University)

Dry granular materials tend to fail by shallow sliding along planar surfaces, inclined at a slope angle which lies a few degrees below the upper ("static") angle of repose (see Type 13, above). However, because of strength homogeneity, the motion of dry granular material changes to shear distortion and the movement becomes flow like. In the absence of pore-pressure changes, the movement tends to be slow, because the difference between the

Fig. 21 The 1999 rock avalanche deposited on a glacier surface, Mt. Munday, British Columbia, Canada (Delaney and Evans 2013). (Topography and ortho-photo courtesy of MacElhanney, Ltd., Vancouver and image courtesy S.G. Evans, University of Waterloo)

maximum and minimum repose angle is small and the potential energy loss is largely compensated by friction. Important geomorphic processes of this type include talus slides and sand slides on the lee slopes of sand dunes (Fig. 22). According to Terzaghi (1957), underwater slides of coarse granular soils, which happen largely under drained conditions and do not develop excess pore-pressure, can also be included in the same category.

20. *Sand/silt/debris* flowslide: Very rapid to extremely rapid flow of sorted or unsorted saturated granular material on moderate slopes, involving excess pore-pressure or liquefaction of material originating from the landslide source. The material may range from loose sand to loose debris (fill or mine waste), loess and silt. Usually originates as a multiple retrogressive failure. May occur subaerially, or under water.

Loose, saturated granular soils can fully or partially liquefy during or after failure and create extremely rapid flowslides. The earliest descriptions of liquefaction flowslides relate to events occurring underwater and involving loose deltaic deposits, as well as hydraulic fills (e.g., Bjerrum 1971; Casagrande 1940; Koppejan et al. 1948; Locat and Lee 2002). The rapidity and long displacement of the underwater flows is often evidenced only indirectly, by sudden removal of large volumes of sediment from the sea floor and generation of surface waves. Bjerrum (1971) described a barge being dragged for tens of metres by its anchor, trapped in an underwater flowslide. McKenna et al. (1992) described an event which took place on the front of the Fraser Delta near Vancouver in 1985. A layer of loose sand, 20 m thick and 75,000 m^2 in area "disappeared" from the delta front during a 3-h period, while a sounding survey was in progress. Despite the rapid displacement of such a large volume, no wave activity was observed in the area, suggesting that the material spontaneously liquefied in a gradual, retrogressive manner and flowed away as a density current.

In all the cases mentioned above, the trigger was spontaneous liquefaction, caused by over-stressing of loose-saturated soil, probably aided by underconsolidation of the rapidly aggrading deltaic sand deposits (Morgenstern 1967). Of course, earthquake liquefaction can produce similar effects on a much larger scale where the conditions allow.

Subaqueous flowslides can enlarge by undrained loading and entrainment of loose substrate, as evidenced by submarine canyons radiating from source areas. Many become diluted and continue moving over distances as great as hundreds of kilometers, in the form of submarine density currents. The most spectacular example is the 1929 Grand Banks, Newfoundland, underwater landslide, triggered by an earthquake with a magnitude of 7.2. A recent review of this event is given by Fine et al. (2005).

Under subaerial conditions, liquefaction-prone material can be isolated in certain horizons, while a large part of the soil profile can be relatively dense, or even unsaturated (Hutchinson 1992b). The most spectacular are flowslides in loess (aeolian silt and fine sand), which is primarily unsaturated and weakly cemented (e.g., Dijkstra et al. 1994). The catastrophic 1983 landslide at Sale Shan in the Gansu Province of Central China is an example (Fig. 23). Here, an accumulation of loess, perched on a slope of landslide-prone Cretaceous shale, collapsed following a period of rainfall and flowed more than 1 km across the valley at extremely high velocity (Zhang et al. 2002). The surface of the flowing mass was dry and coherent. An eyewitness was carried on top of the flow, without injury. Liquefaction of a basal zone of loess, saturated by groundwater perched above the contact with the shale substrate was likely the cause, while overstress of the loess because of bedrock shearing was likely the trigger (Derbyshire et al. 1991).

Similar flowsliding in loess, triggered on a widespread and gigantic scale by the M 7.8 Gansu Earthquake of December, 1920,

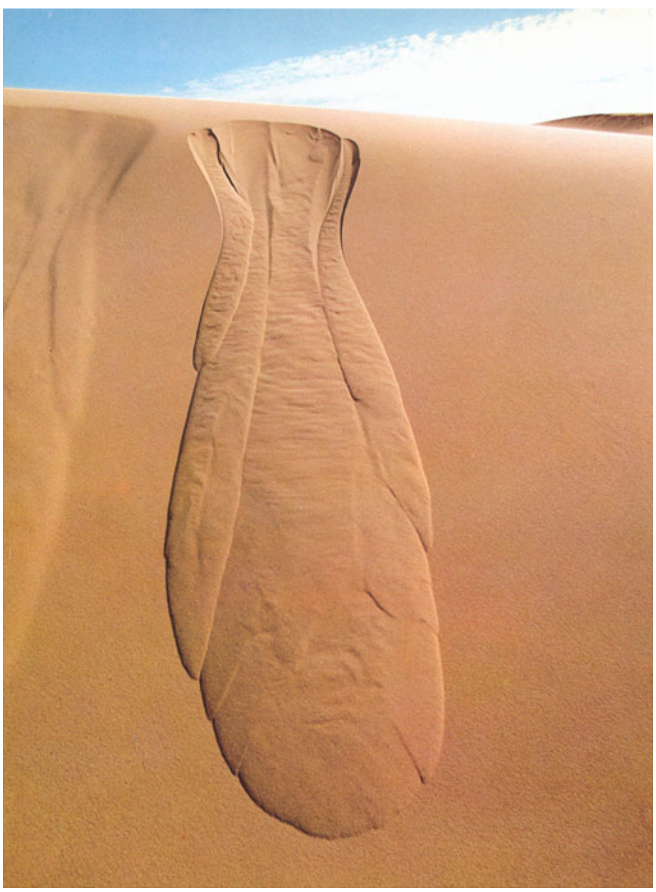

Fig. 22 Dry sand flow on the lee slope of a sand dune, Namib Desert (Courtesy of G.D. Plage)

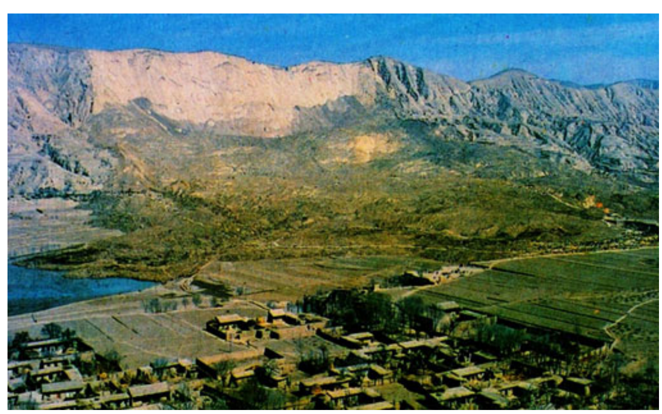

Fig. 23 The Sale Shan flowslide in the loess deposits of the Gansu Province, China, which killed 237 persons in March, 1983 (Courtesy of G. Wang, DPRI, Kyoto University)

was the origin of the most deadly landslide disaster in history, with more than 100,000 fatalities (e.g., Zhang and Wang 2007).

The landslide literature occasionally alludes to the possibility of liquefaction of dry, fine-grained soil by air pressure (e.g., Varnes 1978). An example is one of the several spectacular flowslides in industrial waste, the Jupille fly ash failure, reviewed by Bishop (1973). Considering maximum compression of air as a result of collapse densification of very loose granular material, full or partial liquefaction of flows up to several metres thick may be possible (Hungr 1981, p. 243). The role of water in such landslides is difficult to quantify. The surface may appear dry, but saturated material may be concealed at depth.

The distinction between a flowslide and liquefaction spreading (no. 16 above) is gradational and based largely on judgment. The latter type should be used for landslides that contain a large proportion of laterally displaced solid blocks, rafted on a liquefied base.

The occurrence of liquefaction flowsliding is constrained to a certain, narrowly defined group of "liquefiable" materials, forming a significant part of the source volume. The most common among these are loose sands or gravels under water (or under the water table), loose glacio-fluvial silts or loess with basal saturation, loose man-made fills, mining waste (e.g., Hungr et al. 2002) or mine tailings (Blight 1997) or air fall pyroclastic soils (Picarelli et al. 2008).

However, the range of liquefaction-prone materials may be much wider than what conventional experience suggests. Of special concern are occurrences of flowslides in previously failed soils. For example, the Attachie Slide on the Peace River near Fort St. John, British Columbia, Canada, occurred in a sequence of overconsolidated, insensitive clays and silts, that had previously failed in extensive, slow-moving compound slides, accumulating several tens of meters of displacement (Fletcher et al. 2002). The rapid failure of May, 1973 carried 7 million m^3 of this material over a distance of more than 1 km over an average slope angle of 7.7° in less than 1 min, crossing the floodplain of the Peace River and raising a displacement wave 15 m high on the opposite bank. Evidently, the ductile nature of a portion of the material changed as a result of cracking and softening, following the initial instability.

A similar dramatic change of behavior was demonstrated by the 2005 extremely rapid flowslide at LaConchita, California, with a loss of 10 lives (Jibson 2005). The catastrophic flowslide was derived from the debris of a 1996 slow earthflow at the same location. Apparently, the plastic earthflow material was modified into a brittle, liquefiable mass by weathering (softening) over a period of 9 years. This type of liquefiable material cannot presently be identified by standard geotechnical testing.

Earthquake and spontaneous liquefaction should be clearly separated from the processes of rapid, undrained loading, mixing, and dilution, which play a dominant role in earthflows, debris flows and debris avalanches. During the undrained loading process, a sudden increase of total stress in a saturated (or nearly saturated) soil increases the pore pressure while the effective stress remains at a constant low value (e.g., Hutchinson and Bhandari 1971; Sassa 1985). In contrast, during liquefaction the total stress remains constant, but the effective stress is reduced by structural collapse (e.g., Lefebvre 1995; Eden and Mitchell 1970; Picarelli et al. 2008). Rapid undrained loading can affect all materials, even if less than 100 % saturated and is therefore controlled more by the process than by material character.

A special type of flowslides occur in periglacial regions, where liquefaction occurs in loose fine-grained soils saturated by melting of ground ice. McRoberts and Morgenstern (1974) distinguished shallow "skin flows" and deep-seated "bi-modal flows" or "thaw flows" (Fig. 24). Thawing often begins at river banks, where ice-rich soil is exposed by stream erosion. As new scarps are exposed to surface thawing, extensive retrogressive sliding and flow of liquid debris follow (Wang et al. 2009). Both the retrogression rate and flow velocities are typically low.

21. Sensitive clay flowslide: Very rapid to extremely rapid flow of liquefied sensitive clay, due to remolding during a multiple retrogressive slide failure at, or close to the original water content.

Rapid strength loss because of sudden remolding during failure at natural moisture content, a behavior similar in its effect to liquefaction, also occurs in very sensitive, so called "quick" clays, often leached marine clays in terraces created by isostatic uplift (Fig. 25). As this behavior involves a sudden "phase change" of the material from solid to liquid, triggered by shear failure, the traditional geotechnical term "clay flow slide" has long been applied to such landslides in general (e.g., Meyerhof 1957; Eden and Mitchell 1970, and others).

In a recent summary, Locat et al. (2011) distinguished "multiple-retrogressive slides (clay flows)", "progressive translational (flake) slides," and "lateral spreads" (cf. Hutchinson 1988). The term "sensitive clay flowslide" should be applied to the first and sometimes the second category and is distinguished by nearly complete depletion of the source area due to wide-spread remolding and flow of the material.

22. Debris flow: Very rapid to extremely rapid surging flow of saturated debris in a steep channel. Strong entrainment of material and water from the flow path.

Fig. 24 A flowslide caused by multiple retrogressive failure of ice-rich permafrost (a thaw flow) in the Mackenzie Region, North-west Territories, Canada (Courtesy of B. Wang, Geological Survey of Canada)

Fig. 25 A flowslide in sensitive clay, Lemieux, Ontario, Canada (Courtesy of S.G. Evans, University of Waterloo)

Fig. 26 A debris flow drainage and fan in the Khumbu Valley, Nepal (Photo by O. Hungr)

The term "debris flow" (*murgang* in German, *coulée de débris* in French, *dōseki-ryu* in Japanese, and *selevoii potok* in Russian) is a wide-spread hazardous phenomenon in mountainous terrain. It is distinct from other types of landslides in that it occurs periodically on established paths, usually gullies and first- or second-order drainage channels. Thus, debris flow hazard is specific to a given path and deposition area ("debris fan"). This, and the periodicity of occurrence at the same location, influences the methodology of hazard studies and contrasts with related phenomena, such as debris avalanches (type 25), whose occurrence is not bound to an established path.

Debris flow events often occur simultaneously with floods. The flow may be initiated by a slide, debris avalanche or rock fall from a steep bank, or by spontaneous instability of the steep stream bed. Once soil material begins to move in a steep channel, the bed becomes subject to rapid undrained loading, often so sudden that it could be characterized as impact loading (Sassa 1985). Under such conditions, even coarse material can liquefy, or at least suffer a significant increase in pore-pressure. The bed material will become entrained in a growing surge. As the surge moves downstream, erosion undermines the steep banks and further soil material, as well as organic debris, is added to the flow. The surges travel down the channel on slopes steeper than 10–20° and entrain saturated soil, as well as surface water present in the channel. The bulk of the material involved in a debris flow event usually originates from entrainment from the path, while the volume of the initiating slide is insignificant. The magnitude of debris flows therefore depends primarily on the characteristics of the channel and can be estimated by empirical means (Hungr et al. 2005).

A typical debris flow event is shown in Fig. 26. Here, the initiating instability was a small rock slide, which projected rock fragments directly into the steep upper channels of the small drainage. Surges formed and the flow increased in volume, while travelling through the steep middle region ("gorge") of the channel. Tributary instabilities added to the volume. Finally, the debris flow event deposited on a debris flow fan, having traversed a distance of about 3 km with a vertical drop of 2,000 m.

As a result of channelization, a debris flow surge grows and becomes fronted either by a boulder concentration (e.g., Pierson 1986) (Fig. 27) or a turbulent "head" (Davies 1986) (Fig. 28). Periodic damming and release may occur, contributing to surge growth. A debris flow event may consist of a single surge, or many, as documented in a time-series diagram of discharge at a point in the channel, recorded by Hübl et al. (2009) at Lattenbach, a mountain torrent in the Austrian Alps. Here, 11 surges exceeding 50 m^3/s in peak discharge occurred during an 8-min period, one of which momentarily reached a maximum of 370 m^3/s. Debris flow surges build steep fronts and their peak discharges are magnified by this (Hungr 2000). The peak discharge of the largest surge

Fig. 27 Debris flow surge in the Kamikamihori Valley. (A frame from a video recording courtesy of H. Suwa of the DPRI, Kyoto, Japan)

Fig. 28 The head of a debris flow surge in Jiang-Jia Ravine, Western China, maintained by turbulence (Courtesy of K.M. Scott, US Geological Survey)

Fig. 29 Mudflow deposits, Chilliwack Valley, Canada (Photo by O. Hungr)

involved in a debris flow event may be more than 1 order of magnitude greater than the most extreme hydrological flood (VanDine 1985). This high discharge is responsible for great flow depth, high impact loads and the ability to move large boulders.

A detailed comparison of dynamic behavior of debris surges on a volcano in Java, Indonesia is given by Lavigne and Suwa (2004), who found discharge magnification in debris flow surges there to be relatively modest. A complete definition of a debris flow (or mud flow) should therefore consider two parallel criteria, applied to the largest surge in the series forming an "event": (1) the peak discharge is more than three times greater than that of a major flood flow, or (2) mean solids volume concentration at the surge peak is greater than about 60 % and the water and solid phases are thoroughly mixed.

Many debris fans accumulate material from debris flows, together with debris floods and ordinary fluvial bedload. Symptoms used to distinguish debris flow material from other sediment on a fan include high slope angle of the fan, very large individual particles, coarse levees and boulder trains, signs of impact loading on obstacles, U-shaped eroded channels and, of course, steep, debris-loaded channels upstream.

Debris surges grow in a steep, confined channel, but spread out when the channel exits onto the surface of the debris (colluvial) fan, at typical slopes of 5° to 20°. The frontal boulder accumulation rapidly deposits in the form of levees or abandoned boulder fronts, while the finer and more dilute material continues further downslope. In this fashion, even fully developed debris flow surges eventually convert into debris flood surges described below.

23. Mud flow: Very rapid to extremely rapid surging flow of saturated plastic soil in a steep channel, involving significantly greater water content relative to the source material. Strong entrainment of material and water from the flow path (Plasticity Index>5 %).

In some regions, debris flows transport primarily coarse granular debris, containing only a small proportion of silt and clay. In regions of sedimentary, volcanic and metamorphic rocks and those with deep weathering, the material may contain significant content of fines and be measurably plastic (Bull 1964). Such soil drains more slowly and remains longer in a liquid condition, leading to longer travel and lower slope angles in the deposition area (Fig. 29). Because of these special characteristics, a distinct term "mud flow" is useful. The boundary between debris flow and mud flow is gradational. Here, the Plasticity Index of the material is suggested as the controlling parameter, although it may be desirable in the future to find an index using also the percentage of silt and finer grain sizes.

Mud flows occur on a very large scale on stratovolcanos, where they exploit the abundance of fine-grained deposits of pyroclastic material and ash (e.g., Vallance 2005). The supply of large volumes of water needed for mud flow generation may be from precipitation, from melting of summit ice or snow by volcanic heat, or from glacier outburst flooding (jökulhlaup). In November, 1985, gigantic mud flows were triggered by the melting of the summit ice cap of the 5,389-m Nevado del Ruiz in Colombia, during a minor eruption (Pierson et al. 1990). They destroyed the town of Armero and caused 23,000 fatalities, becoming one of the deadliest landslide disasters of recent time.

Large mud flows or debris flows from volcanic sources are often referred to by the Indonesian term "lahars" and can occur during eruptions ("hot lahars") or during periods of high surface water runoff while the volcano is dormant ("cold lahars").

24. Debris flood: Very rapid flow of water, heavily charged with debris, in a steep channel. Peak discharge comparable to that of a water flood.

During extreme flooding in steep channels, the stream bed may be destabilized causing massive movement of sediment. Such sediment movement (sometimes referred to as "live bed" or "carpet flow" by hydraulicians) can reach transport rates far exceeding normal bed load movement through rolling and saltation. However, the movement still relies on the tractive forces of water. Large quantity of sediment may be transported to the debris fan, but the peak discharge remains in the same order as that of a flood, even if magnified by a "bulking rate" of up to 2–3, approximately (Costa 1984). Unlike a debris flow, a debris flood usually does not develop high impact forces and potential damage to structures is limited.

However, this depends on the size of the drainage where the debris flood originated and the origin of the water discharge. While debris flows are limited to steep drainages of less than a few square kilometers, debris floods can occur in much larger watersheds, with correspondingly greater hydrologic flood discharges, often called "flash floods" and are magnified by heavy sediment loads. For example, during the catastrophic flood and landslide disaster of the Serrana Region of Brazil, in January, 2011, rivers with drainage areas exceeding 50 km^2 experienced extreme flooding. Figure 30 shows a scene along the path of the flooding, near the apex of an alluvial fan, with a slope of 4°. The boulder in the centre of the photo, one of many spread over the fan, appears to have been rolled by the water flow, which was nearly 2 m deep at this location. The weak concrete structure on the right was only slightly damaged by the same discharge, showing that the sediment concentration in the flow was limited.

Extremely large and damaging debris floods occur in small or medium size watersheds in mountainous terrain due to outburst and sudden drainage of moraine-dammed pro-glacial lakes. Such glacial lake outburst floods ("GLOFs") have periodically caused serious damage in glaciated mountains of Nepal (ICIMOD 2011), North America (Evans and Clague 1994), Russia (Chernomorec 2005), and South America. Some GLOFs reach distances of tens of kilometers from the source outburst and may attain the character of true debris flows in their largest surges (A. Strom, Russian Academy of Science, personal communication, 2013).

Considering the water drag, debris flood deposits extend further downslope than debris flows and deposit on smaller slope angles (often less than 5°). There is a continuum between "clear" water floods and debris flows, as recognized by Stini (1910), Hutchinson (1992a) and many others. The distinction between debris floods and debris flow surges is of great practical importance due to their different damage potential and also because of the widely different strategies that must be used to design protective structures.

25. Debris avalanche: Very rapid to extremely rapid shallow flow of partially or fully saturated debris on a steep slope, without confinement in an established channel. Occurs at all scales.

In coining the term debris avalanche, Sharpe (1938) compared the morphology of such shallow slides on steep slopes to that of snow avalanches. In contrast to a debris flow, a debris avalanche is a unique event that can be found anywhere on steep slopes. This difference is decisive for the selection of methodology during hazard studies. In many cases, debris avalanches enter established channels, de-stabilize channel infills, and become debris flows.

Debris avalanches initiate as debris slides (type 13) and are associated with failures of residual soil, colluvial, pyroclastic, or organic veneer (Fig. 31). In some cases, failure of thicker accumulations of granular material on steep slopes, such as deep pockets of residual soil or artificial loose fills, may initiate large debris avalanches. In such cases, it is difficult to make distinction between debris avalanches and flowslides (type 20). The source volumes of debris avalanches may contain liquefiable material (Picarelli et al. 2008). Cohesion loss, spontaneous liquefaction, and undrained loading can all occur simultaneously in a landslide on a steep slope. However, it is suggested that the term flowslide be reserved for failures where spontaneous or earthquake liquefaction is clearly the dominant mechanism.

The rapid undrained loading process also allows debris avalanches to be triggered by impact from rock fall or rock slide on soil-covered slopes (e.g., Lacerda 2007).

Once initiated, rapid undrained loading continues progressively as material moves down the slope. In this way, an initial landslide of a few tens of m^3 can strip material from a large segment of the slope, entraining many thousands m^3. Prediction of potential debris avalanche magnitude (volume) therefore requires both the estimate of the thickness of entrainable layer and the plan

Fig. 30 Debris flood damage during the January, 2010 flooding at Teresopolis in the Serrana Region, Brazil. The large boulder was rolled by a 2-m deep flow, but the concrete structure on the right was only partly damaged (Photo by O. Hungr)

Fig. 31 A debris avalanche in sandy colluvium, Jasper national Park, Canada. The debris deposited on top of snow avalanche deposits which came from the same source area the during the preceding winter. Note sitting person on the right. (Photo by O. Hungr)

dimensions of the avalanche path. The paths widen downslope, as the undrained loading destabilizes an increasing width of the slope segment (Fig. 32). Guadagno et al. (2005) defined "apex angle" as the angle of widening of the path for debris avalanches from the 1998 Sarno disaster in the Campania Region of Italy and found it to vary between 5° and 50°, depending on the depth of the pyroclastic veneer and the slope angle.

Some of the most catastrophic regional landslide disasters occur as clusters of debris avalanches and debris flows during heavy rainstorms or earthquakes (Fig. 33). The 1999 Vargas State disaster of northern Venezuela, for example, caused the total displacement of some 100,000 m^3 of material per square kilometers, as thousands of debris avalanches descended from steep slopes and the liquid debris concentrated in gigantic debris flows and floods. Approximately 15,000 persons lost their lives (Larsen and Wieczorek 2006). It is of interest that the same area was subject to a similar disaster in 1955, although sparse population of the hazard areas at that time prevented the appalling loss of life.

Numerous clusters of debris avalanching have been experienced in New Zealand during cyclonic storms, covering as much as 30 % of an area of steep slopes by landslide scars (Crozier 2005). The most recent debris avalanche cluster of global significance was the January, 2011 disaster in the Serrana Region of Brazil, where over 3,500 landslides occurred on steep slopes during a period of 3 days, together with debris flows in small, steep drainages and debris floods in larger streams (Fig. 33). Over 1,500 fatalities occurred in the populated rural region (Avelar et al. 2011; Coelho Netto et al. 2011). A disaster of similar magnitude occurred in the same part of Brazil in 1967 (Schuster et al. 2002).

Fig. 32 Several of the catastrophic debris avalanches of May, 1998 in Siano, Italy, illustrating the characteristic widening of debris avalanche scars on steep slopes (Vertical airphoto published by licence no. 2347-02/December/2002 of Regione Campania. Image courtesy of Dr. P. Revellino, University of Sannio, Benevento)

Fig. 33 A cluster of debris avalanches and debris flows of January, 2010 in the Serrana Region of Brazil (Courtesy of A.L. Coelho-Netto, Federal University of Rio de Janeiro)

Debris avalanches move at extremely high velocities. In 2011, video footage of debris avalanches in Seoul, South Korea was released on the Worldwide Web, showing debris moving down heavily forested slopes and impacting urban roads and buildings at speeds of more than 20 m/s (70 km/h).

Some large individual debris avalanches can be particularly dangerous, if they occur in exceptional circumstances. An example is the July, 2012 debris avalanche at Johnson Landing, a small community in the interior of British Columbia, Canada. The source was an accumulation of glacial deposits of sand and silt, which had for a number of years been subject to shallow rotational and translational sliding. During a rainy period combined with snowmelt in early July, 2012, 300,000 m^3 of this previously disturbed material failed and flowed down a small creek channel at extremely high speed. The event spilled out of the established channel covered the surface of a terrace where no landslide debris had deposited before. Four lives and several houses were lost. Such hazard scenarios depend on unique sets of circumstances and are extremely difficult to anticipate and prevent.

26. Earthflow: Rapid or slower, intermittent flow-like movement of plastic, clayey soil, facilitated by a combination of sliding along multiple discrete shear surfaces, and internal shear strains. Long periods of relative dormancy alternate with more rapid "surges".

Earthflows occur in plastic, disturbed, and mixed soils, whose consistency lies close to the Plastic Limit (Keefer and Johnson 1983). Such material deforms easily, but is essentially ductile and does not significantly lose strength during deformation. As a result, earthflows move slowly and intermittently. The intermittent character of earthflow motion is especially pronounced in arid climates. The highest localized earthflow surge speed documented in the literature is 0.13 m/s (Hutchinson et al. 1974), however, typical movement velocities are measured in meters per hour during surges (Picarelli et al. 2005) and meters per year in general (e.g., Bovis 1985). Earthflows occur on slopes typically inclined at less than about 12° and vary in length from a few tens of meters to 6 km (Varnes and Savage 1996).

Many earthflow tongues remain in a dormant state for many decades, allowing roads and buildings to be built on them. For example, 2,000 villages in the Northern Italian region of Emilia Romagna are built on earthflow terrain and periodic damage takes place (Bertolini et al. 2005). The Thistle earthflow in Utah, USA, had been dormant and unrecognized since pre-historic time. A natural surging episode in 1985, however, blocked the Spanish Forks River and necessitated the construction of an emergency spillway, bypass tunnel, and relocation of a highway and railroad, making this the costliest landslide in the history of the USA (Schuster and Highland 2001).

As defined by Keefer and Johnson (1983), earthflows constitute a transporting agent between a source slide area and an eroding toe. The source can be one or a series of rotational or compound slides or a weathering and eroding steep face in weak rock. The material becomes softened without absorbing excessive moisture, so it remains in a plastic state (Liquidity Index<0.5). The body of the earthflow slowly deforms, or fails along multiple shear surfaces, producing lobate, flow-like morphology. Acceleration ("surging") occurs when the source slide becomes destabilized, usually by a temporary increase in pore-pressure. As material in any part of the earthflow tongue accelerates, it over-rides or compresses soil masses downslope, increasing pore pressure through undrained loading (Hutchinson and Bhandari 1971). In this way, a kinematic wave propagates through the soil mass, to advance the toe into a stream, a water body, or another erosional sink. An earthflow undergoing a surge is shown in Fig. 34. A photographic analysis documenting the detailed history of surging of this earthflow over a period of 56 years has been published by Guerriero et al. 2013.

The kinematics of the flow-like motion varies. During slow, steady motion phases, the deformation may be concentrated on the main shear surface, analogous to a translational slide (hence the English term "mudslides," Hutchinson 1988). During surges, numerous internal shears ("imbricate thrusts") develop and combine with distributed internal strains of the plastic mass to generate flow-like morphology (e.g., Bertolini et al. 2005). Hazard assessment for earthflow areas requires the identification and prediction of areas subject to surging re-activation, producing local velocities in the slow to rapid range.

Fig. 34 An earthflow surge initiated by a planar slide in the Campania Region, Italy (Courtesy of F. Guadagno University of Sannio, Benevento)

In certain areas and geological units, earthflows in varying states of activity cover a large percentage of sloping ground, forming extensive complexes. As an example, gentle dip slopes in tectonized clays of the Campania Region in southern Italy comprise earthflow complexes over nearly 50 % of their total surface area (Revellino et al. 2010).

27. Peat flow: Rapid flow of liquefied peat, caused by an undrained failure

Peat is a light, organic material with varying degrees of fibrous texture. The presence of organic fibres and mineral grains gives peat a fairly high drained friction angle, often in excess of 30°. The extreme compressibility and high water content, however, make the material susceptible to dramatic weakening during undrained loading. Once an initial movement begins and peat layers are subjected to compression, extreme loss of strength occurs, followed by flow. The initial movement is often caused by human activity, especially rapid placement of artificial fill on organic substrates. However, there is also a range of natural processes by which the margins of over-saturated peat accumulations fail ("bog bursts"). Some of these have attained speeds sufficient to cause fatalities. A detailed review and classification of landslides in peat has recently been published by Dykes and Warburton (2007).

Slope deformation

28. Mountain slope deformation: Large-scale gravitational deformation of steep, high mountain slopes, manifested by scarps, benches, cracks, trenches and bulges, but lacking a fully defined rupture surface. Extremely slow or unmeasurable movement rates.

Highly stressed rock masses forming mountain slopes, with vertical relief of the order of 1 km or more, can be subject to visible deformation, evidenced by topographic surface features such as scarps, scarps, benches, cracks, trenches, and bulges. Large-scale deformation of mountain slopes was first described by Heim (1932) who considered such phenomena as precursory signs of impending slope failure. As noted by Zischinsky (1969), so-called slope sagging ("sackung") is ubiquitous on mountain slopes. The term "slope creep" is also sometimes used. Gravitational slope deformation features are often misidentified as tectonic fault displacements.

In materials science, the term "creep" designates time-dependent deformation under constant stress. Its application to landslides is ambiguous. True creep does occur in geomaterials, especially organic soils (secondary consolidation), ice-rich frozen soils, salt and gypsum, or soils or rocks losing strength by structural deterioration just prior to failure (undergoing tertiary creep). In most cases, however, slow deformation of soils and rocks that is commonly referred to as creep occurs in fact under fluctuating (cyclic) effective stress levels.

There is a wide variety of styles of deformation, depending on the structure of the rock masses, relative to the slope orientation. Hutchinson (1988) offered a classification based on kinematics, which largely parallels the types of sliding and toppling movements. Thus, R-style of deformation resembles the early stages of rotational sliding; CL and CB styles correspond to listric or bi-linear compound sliding respectively and T-style results from toppling. Both single-

Fig. 35 Slope deformation features on Mission Ridge, Southern B.C., Canada. (Courtesy of C. Esposito, La Sapienza, University of Rome)

and double-sided ("ridgetop spreading") examples occur. Figure 35 depicts a double-sided CB-style deformation of a ridge near Lillooet, in British Columbia.

A variety of deformation features was described in the Slovak Carpathians by Nemčok (1982). As shown in Fig. 36, Nemčok recognized temporal development of deformation, starting with subdued initial stages and progressing to advanced and prominent features. He suggested that the final stages of deformation involved the development of discrete rupture surfaces and sliding failure. Cases exist where slopes in various stages of deformation did produce large failures (e.g., Heim 1932). However, a much larger number of cases can be seen where deformed slopes remain stable, at least within historical time. The time rate of development is very slow, so that no record exists documenting the full development of slope deformation features from initial stages to failure. This poses a difficult practical problem for professionals charged with the assessment of hazards on valley floors beneath deformed slopes, which include a large extent of developed areas.

Rates of movement involved in slope deformation are slow and often unmeasurable. Moser (1996) presented movement rates ranging from a few centimeters to several meters per year and commented that the rates are highly variable in both time and space. Many slopes move at rates clearly controlled by groundwater levels in the slope, accelerating during spring thaw periods. Some discrete movements occur during earthquakes (A. Strom, personal communication, 2011). Prominent deformation features often appear behind the main scarp of large rock slides (e.g., Froese et al. 2009). Increased deformations are also observed in slopes influenced by subsidence from mining operations. The 1903 failure of the Frank Slide in Alberta, Canada, was preceded by deformations caused by stoping in a coal mine, situated at the toe of the slope (Benko and Stead 1998). Prominent mountain slope deformations in the Italian Apennines have been ascribed to karstic subsidence (Discenza et al. 2011).

29. Rock slope deformation: Deep-seated slow to extremely slow deformation of valley or hill slopes. Sagging of slope crests and development of cracks or faults, without a well-defined rupture surface. Extremely slow movement rates.

Even relatively small slopes made of weak claystones and marls deform, especially if capped by a massive layer of stronger rock. The widespread phenomenon of "cambering," "bulging," and "gulls" along the crests of British river valleys has been discussed in many publications. Hutchinson (1991) suggests that many of these features may have formed due to the flow of ice-rich clayey rocks during the periglacial conditions in Early Holocene.

Pleistocene glacial meltwater-enlarged valleys of western Canada are subject to similar phenomena, termed "valley rebound" (Matheson and Thomson 1973). The symptoms of such deformations include sagging of slope crests, separation and toppling of surficial blocks, formation of tension features behind the crests of slopes, flexural pre-shearing of bedding planes and uplift of valley floors. Generally, these deformations do not lead to slope failure, although they may facilitate the development of compound rock slides by pre-shearing weak horizons in the Cretaceous shale rock.

The difference between rock slope deformation (type 29) and mountain slope deformation (type 28) is scale, with the former involving slopes only a few tens or hundreds of meters high and weak rock. Many rock slope deformation cases involve break-up of strong cap rock overlying a weak clayey unit and should be classified as rock slope spreading (Type 15).

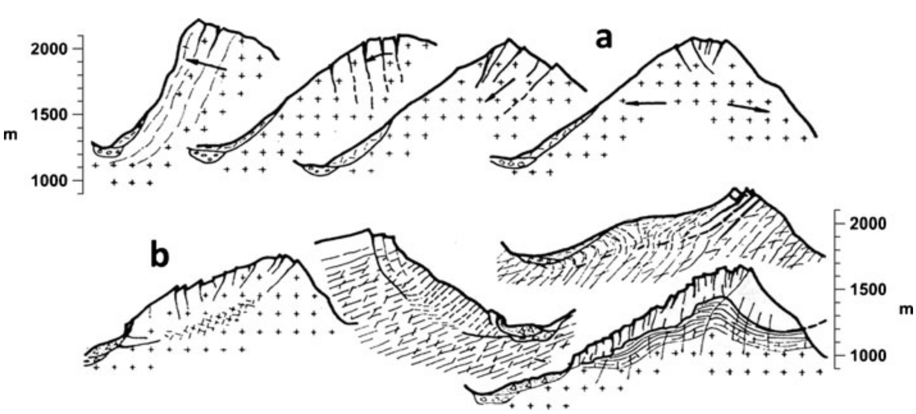

Fig. 36 Examples of mountain slope deformation from the Slovak Carpathians. **a** Initial and **b** advanced stages (Nemčok 1982)

Fig. 37 Shallow planar slide-earthflow, Campania Region, Italy (Photo by O. Hungr)

30. Soil slope deformation: Deep-seated, slow to extremely slow deformation of valley or hill slopes formed of (usually cohesive) soils. Often present in permafrost slopes with high ice content.

Slow deformations also occur in soils adjacent to steep slopes. On-going deformations of the crests of excavated slopes are well known to geotechnical engineers and often necessitate support measures to prevent damage to adjacent roadways and structures. Even larger deformations occur at the crests of rapidly built fill slopes. For example, large mine waste piles built by end dumping on foundations of steep terrain can deform by several tens of metres. The deformation rates can reach more than 1 m/day and result from a combination of volumetric consolidation and plastic yielding of the waste and foundation. They sometimes, but not always culminate in extremely rapid flowslides (Hungr et al. 2002).

Under natural conditions, actively eroding slopes along shorelines and riverbanks also demonstrate gradual deformation. The displacements are usually small, except in cases where the soil is close to sliding failure. Stable, long-term deformations can occur where permafrost with significant ice content exists within the slope (Savigny and Morgenstern 1986).

31. Soil creep: Extremely slow movement of surficial soil layers on a slope (typically less than 1 m deep), as a result of climate-driven cyclical volume changes (wetting and drying, frost heave).

Table 5 Summary of the proposed new version of the Varnes classification system. The words in italics are placeholders (use only one)

Type of movement	Rock	Soil
Fall	1. *Rock/ice* fall[a]	2. *Boulder/debris/silt* fall[a]
Topple	3. Rock block topple[a]	5. *Gravel/sand/silt* topple[a]
	4. Rock flexural topple	
Slide	6. Rock rotational slide	11. *Clay/silt* rotational slide
	7. Rock planar slide[a]	12. *Clay/silt* planar slide
	8. Rock wedge slide[a]	13. *Gravel/sand/debris* slide[a]
	9. Rock compound slide	14. *Clay/silt* compound slide
	10. Rock irregular slide[a]	
Spread	15. Rock slope spread	16. *Sand/silt* liquefaction spread[a]
		17. Sensitive clay spread[a]
Flow	18. *Rock/ice* avalanche[a]	19. *Sand/silt/debris* dry flow
		20. *Sand/silt/debris* flowslide[a]
		21. Sensitive clay flowslide[a]
		22. Debris flow[a]
		23. Mud flow[a]
		24. Debris flood
		25. Debris avalanche[a]
		26. Earthflow
		27. Peat flow
Slope deformation	28. Mountain slope deformation	30. Soil slope deformation
	29. Rock slope deformation	31. Soil creep
		32. Solifluction

For formal definitions of the landslide types, see text of the paper.

[a] Movement types that usually reach extremely rapid velocities as defined by Cruden and Varnes (1996). The other landslide types are most often (but not always) extremely slow to very rapid

Soil and weak rock layers within one metre, approximately, of the ground surface are subject to cyclical volume changes due to swelling and shrinkage with moisture changes, freezing and thawing, and plant and animal activity. As explained in textbooks of geomorphology, volumetric expansion acts normal to the sloping ground surface while, during shrinkage, the material moves vertically down under gravity. The result is a net downslope movement, termed soil creep (e.g., Sharpe 1938). The rates of movement are extremely slow (0.5 to 10 mm/year as compiled by Saunders and Young 1983), but, over long periods of time, most steep slopes become mantled by loose, displaced, and mixed layer of colluvium. The surficial layer loosened and mixed by soil creep is often the primary source of shallow soil slides and debris avalanches.

This cyclic phenomenon has nothing in common with the mechanistic meaning of the term "creep" and is used here only out of respect to the long-established use of the term in the geomorphology literature. Because of its established status, the term "soil creep" does not have a suitable alternative.

32. Solifluction: Very slow but intensive shallow soil creep involving the active layer in Alpine or polar permafrost. Forms characteristic solifluction lobes.

Soil creep is intensified approximately tenfold by the presence of seasonal ground ice in the surficial soil under alpine or periglacial conditions. One reason is that ice is capable of true creep, i.e., deformation under constant stress. In addition, during the thawing season, the active layer overlying the impervious permafrost table becomes charged with water and normal movements due to volume changes are added to by episodic sliding deformation.

Complex landslides and secondary effects

Although the proposed classification system does not contain a separate class of complex landslides, it is sometimes necessary to use two type names to describe a case, where a unique type cannot be assigned. As stated earlier, the need for such composite terminology should be decided by the user of the classification. This section provides some examples.

Rotational (or planar or compound) slide-earthflow ("slump earthflow" of Varnes 1978) is a relatively small landslide, where a sliding failure provides the source to an earthflow of limited extent (Fig. 37). As both stages of the slope movement are of similar extent and significance, it is difficult to place the event into either of the two separate categories (slide and flow) and a composite name is useful.

Both rock falls and rock slides may impact saturated talus or other soil deposits and mobilize debris avalanches or debris flows. Hungr and Evans (2004b) documented several cases where rock slides mobilized colluvial debris avalanches of volume comparable to the size of the initial instability and proposed the term rock slide-debris avalanche for such events. Deline et al. (2011) described a rock slide-debris flow event from the southern Mont Blanc area, Italy. Here, an irregular rock slide of 500,000 m^3 mobilized a comparable volume of colluvium from an apron at the foot of the rock slope. The liquefied debris entered two established debris flow channels, entrained additional material, and snow and travelled to deposit on debris fans in the trunk valley, 2 km distant from the landslide source.

The development of a two-stage rock slide-debris avalanche was documented in a video footage at Preonzo, Canton Valais, in southern Switzerland. Here, a part of actively deforming rock slope failed in an irregular manner (Fig. 10) and covered a steep talus slope by new rock avalanche deposits. After a delay of some minutes, the talus slope began to move as a translational slide, reached rapid velocity and deposited at the foot of the slope (S. Löw, ETH, Zurich, personal communication, 2012). A suitable composite name for such an event would be "irregular rock slide-debris slide". The limited velocity of the debris slide stage indicates an absence of liquefaction, although the talus was clearly mobilized by an increase in pore pressure, due to undrained loading.

The most important secondary effects, which form classes of natural hazards in their own right, include landslide dams and waves generated in reservoirs by landslide impact. A detailed review of landslide dam hazards has been provided in a special volume edited by Schuster (1986) and more recently by Evans et al. (2011). The problem of landslide-generated waves on reservoirs was reviewed comprehensively by Slingerland and Voight (1979). Prediction of landslide-generated waves often relies on parallel application of physical and analytical models (e.g., Zweifel et al. 2007).

Summary

A summary of the proposed landslide type classes appears in Table 5. There are now 32 landslide-type keywords, compared with 29 used in the 1978 Varnes Classification, thus the system remains simple. Ambiguities will always remain, especially where types grade from one to another, without a clear-cut boundary. However, the proposed revision of the Varnes classification meets the objectives laid out at the beginning of this article and the authors hope the classification may be found useful by researchers and practitioners.

Acknowledgments

The initiative aimed at revising the landslide classification was launched in 2006 by the former Chair of JTC-1, Professor Robin Fell. He took an active part in early discussions that led to the drafting of this paper. The latest draft of the paper benefitted from constructive suggestions provided by M. Church, J. Corominas, D. Cruden, A.P. Dykes, H. Einstein, P. Flentje, F. Guadagno, W. Lacerda, V. Merrien-Soukatcheff, R. Poisel, K. Sassa, G. Scarascia-Mugnozza, S. Löw, A. Strom, K. Turner, H.N. Wong, and Y.P Yin. An Associate Editor and two reviewers provided further constructive comments.

References

Antoine P, Camporota A, Giraud A, Rochet L (1987) La menace d'écroulement aux Ruines de Séchillienne. Bulletin, Liaison Laboratoire des Pontes et Causées 150:55–64

Avelar AS, Coelho Netto AL, Lacerda WL, Becker LB, Mendonça MB (2011) Mechanisms of the recent catastrophic landslides in the mountainous range of Rio de Janeiro, Brazil. Procs., 2nd World Lansdlides Forum, Rome

Baltzer A (1875) Über bergstürze in den Alpen. Verlag der Schabelitz'schen buchhandlung (C. Schmidt), Zurich, 50p

Bates RL, Jackson JA (eds) (1984) Glossary of geology. American Geological Institute, Falls Church, Virginia, 788p
Benko B, Stead D (1998) The frank slide: a reexamination of the failure mechanism. Canadian Geotech J 35:299–311
Bertolini G, Guida M, Pizziolo M (2005) Landslides in Emilia-Romagna region (Italy): strategies for hazard assessment and risk management. Landslides 2:302–312
Bishop AW (1973) The stability of tips and spoil heaps. Q J Eng Geol 6:335–376
Bjerrum L (1971) Subaqueous slope failures in Norwegian fjords. In: Proceedings of the First International Conference on Port and Ocean Engineering Under Arctic Conditions, 1:24–47
Blight GE (1997) Destructive mudflows as a consequence of tailings dyke failures: geotechnical engineering. In: Proceedings, Institution of Civil Engineers 125:9–18
Bourrier F, Dorren L, Hungr O (2013) The use of ballistic trajectory and granular flow models in predicting rockfall propagation. Earth Surface Processes and Landforms 38:435–440
Bovis MJ (1985) Earthflows in the interior plateau: southwest B.C. Canadian Geotech J 22:313–334
Bozzano F, Lenti L, Martino S, Paciello A, Scarascia MG (2008) Self-excitation process due to local seismic amplification responsible for the 31st October 2002 reactivation of the Salcito landslide (Italy). J Geophys Res 113, B10312
Bull WB (1964) Alluvial fans and near-surface subsidence in Western Fresno County, California, U.S. Geological Survey Professional Paper 437-A. U.S. Geological Survey, Denver, CO
Cannon SH (1993) An empirical model to predict debris flow travel distance. In: Shem, H.W. and Wen, F. (eds) Proceedings, ASCE Hydraulic Engineering, '93. Pp. 1768–1773
Cannon SH, Gartner JE (2005) Wildfire related debris flow from a hazards perspective. Chapter 15. In: Jacob, M., and Hungr, O. (eds) Debris-flow hazards and related phenomena: Springer, Berlin. pp. 321–344
Canuti P, Casagli N, Garzonio CA, Vannocci P (1990) Lateral spreads and landslide hazards in the Northern Apennine: the example of the Mt. Fumaiolo (Emilia-Romagna) and Chiusi della Verna (Tuscany). In: Proceedings 6th Congr. IAEG, Amsterdam, 3:1525–1533
Cascini L, Cuomo S, Guida D (2008) Typical source areas of May 1998 flow-like mass movements in the Campania region, Southern Italy. Eng Geol 96:107–125
Casagrande A (1940) Characteristics of cohesionless soils affecting the stability of slopes and earth fills. Contributions to soil mechanics, 1925 to 1940. Boston Society of Civil Engineers, pp. 257–276
Chernomorec CC (2005) Selevoye ochagi do i posle katastrof (Mud flows during and after catastrophic outbursts) Nauchnii Mir (World of Science), Moscow, 184p (in Russian)
Chigira M, Kiho K (1994) Deep-seated rockslide-avalanches preceded by mass rock creep of sedimentary rocks in the Akaishi Mountains, central Japan. Eng Geol 38:221–230
Coelho Netto AL, Sato AM, Avelar AS, Vianna LGG, Araujo IS, Croix D, Lima P, Silva AP, Pereira R (2011) The extreme landslide disater in Brazil. In: Proc. 2nd World Lansdlides Forum, Rome
Costa JE (1984) Physical geomorphology of debris flows. In: Costa JE, Fleisher PJ (eds) Developments and Applications in Geomorphology: Springer, Berlin, pp. 268–317
Crosta GB, Frattini P, Fugazza F, Caluzzi L (2005) Cost-benefit analysis for debris avalanche risk management. In: Hungr O, Fell R, Couture R, Eberhardt E (eds) Landslide risk management. Proceedings Vancouver Conference. Taylor and Francis Group, London, pp 517–524
Crozier MJ (2005) Multiple-occurrence regional landslide events in New Zealand. Hazard management issues. Landslides 2:247–256
Cruden DM (1989) Limits to common toppling. Canadian Geotech J 26:737–742
Cruden DM, Antoine P (1984) The slide from Mt. Granier, Isére and Savoie, France on Nov. 24, 1248. In: Proc. 4th. International Symposium on Landslides, Toronto, vol. 1, pp. 475–481
Cruden DM, Hu XQ (1992) Rock mass movements across bedding in Kananaskis country, Alberta. Canadian Geotech J 29:675–685
Cruden DM, Varnes DJ (1996) Landslide types and processes. In: Turner AK, Schuster RL (eds) Landslides investigation and mitigation. Transportation research board, US National Research Council. Special Report 247, Washington, DC, Chapter 3, pp. 36–75
Dai FC, Lee CF (2003) Landslide characteristics and slope instability modeling using GIS, Lantau Island. Hong Kong. Geomorphology 42:213–228
D'Alessandro GR, Berti M, Urbani A, Tecca PR (2002) Geomorphology, stability analyses and the stabilization works on the Montepiano travertinous cliff (Central Italy). In: Allison RJ (ed) Applied geomorphology—theory and practice. Wiley, New York, pp. 21–38
Davies TRH (1986) Large debris flows: a macroviscous phenomena. Acta Mechanica 63:161–178
Delaney KB, Evans SG (2013) The 1997 mount Munday landslide, British Columbia; behaviour, dynamic analysis, and fragmentation of a rock avalanche on a glacier surface. Landslides 13 (in press)
Deline P, Alberto W, Broccolato D, Hungr O, Noetzli J, Ravanel L, Tamburini A (2011) The December 2008 Crammont rock avalanche, Mont Blanc massif area, Italy. Nat Hazards Earth Syst Sci 11:3307–3318
Derbyshire E, Wang J, Jin Z, Billard A, Egles Y, Kasser M, Jones DKC, Muxart T, Owen L (1991) Landslides in the Gansu Loess of China. Catena, Cremlingen, Supplement 20, p. 119–145
Dijkstra TA, Rogers CDF, Smalley IJ, Derbyshire E, Li YJ, Meng XM (1994) The loess of north-central China: geotechnical properties and their relation to slope stability. Eng Geol 36:153–171
Dikau R, Brunsden D, Schrott L, Ibsen M-L (Editors) (1996) Landslide recognition: Identification, movement, and causes. Wiley, New York, 1996, 210p
Discenza ME, Esposito C, Martino S, Petitta M, Prestininzi A, Scarascia Mugnozza G (2011) The gravitational slope deformation of Mt. Rocchetta ridge (central Apennines, Italy): geological-evolutionary model and numerical analysis. Bulletin of Eng Geol Environ 70:559–575
Dykes AP, Warburton J (2007) Mass movements in peat: a formal classification scheme. Geomorphology 86:73–93
Eberhardt E (2008) Twenty-Ninth Canadian Geotechnical Colloquium: the role of advanced numerical methods and geotechnical field measurements in understanding complex deep-seated rock slope failure mechanisms. Canadian Geotech J 45:484–510
Eden WJ, Mitchell RJ (1970) The mechanics of landslides in Leda Clay. Canadian Geotech J 7:285–296
ERM-Hong Kong Ltd (1998) Feasibility study for QRA of boulder fall hazards in Hong Kong (GEO Report No. 80). Geotechnical Engineering Office, Hong Kong Government
Evans SG, Scarascia Mugnozza G, Strom A, Hermanns RL (2006) Landslides from massive rock slope failure. Proceedings, Celano Workshop, NATO Science Series, Series VI, 49:53–73
Evans SG, Hungr O (1993) The assessment of rockfall hazards at the base of talus slopes. Canadian Geotech J 30:620–636
Evans SG, Clague JJ (1994) Recent climatic change and catastrophic geomorphic processes in mountain environments. Geomorphology 10:107–128
Evans SG, Tutubalina OV, Drobyshev VN, Chernomorets SS, McDougall S, Petrakov DA, Hungr O (2009) Catastrophic detachment and high-velocity long-runout flow of Kolka Glacier, Caucasus Mountains, Russia in 2002. Geomorphology 105:314–321
Evans SG, Delaney KB, Hermanns RL, Strom A, Scarascia Mugnozza G (2011) Formation and behaviour of natural and artificial rockslide dams. In: Evans SG, Hermanns RL, Strom A, Scarascia Mugnozza G (eds) Natural and artificial rockslide dams. Springer, Berlin, pp 1–76
Fine IV, Rabinovich AB, Bornhold BD, Thomson RE, Kulikov EA (2005) The Grand Banks landslide-generated tsunami of November 18, 1929: preliminary analysis and numerical modeling. Marine Geology 215:45–57
Fletcher L, Hungr O, Evans SG (2002) Contrasting failure behaviour of two large landslides in clay and silt. Canadian Geotech J 39:46–62
Follacci JP (1987) Les mouvements du versant de la Clapiére a Saint Étienne de Tinée (Alpes Maritimes). In: Bull. Liaison Laboratoire des Ponts et Chausées, Paris, 151: 39–54 (in French)
Forlati F, Lancellotta R, Scavia C, Simeoni L (1998) Swelling processes in sliding marly layers in the Langhe region, Italy. In: Evangelista E, Picarelli L (eds) The geotechnics of hard soils-soft rocks. Balkema, Rotterdam, pp 1089–1099
Froese CR, Moreno F, Jaboyedoff M, Cruden DM (2009) 25 years of monitoring on Southe Peak, Turtle Mountain: understanding the hazard. Canadian Geotech J 46:256–269
Gerath RF, Hungr O (1993) Landslide terrain, Scatter River valley, north-eastern British Columbia. Geoscience Canada 10:30–32
Goguel J, Pachoud A (1972) Geology and dynamics of the rockfall of the Granier Range which occurred in November 1248. Bulletin, Bureau de Récherches Geologiques et Miniéres, Hydrogeologie, Lyon, 1:29–38
Goguel J, Pachoud A (1981) Les mouvements de terrain du versant sud du Massif de Platé, Haute Savoie, France. Bull. Liaison Laboratoire des Pontes et Chausées, Spécial X:15-25
Goodman RE, Bray JW (1976) Toppling of rock slopes. Procs., ASCE specialty conference on rock engineering for foundation and Slopes, Boulder, Colo., Vol.2
Guadagno FM, Forte R, Revellino P, Fiorillo F, Focareta M (2005) Some aspects of the initiation of debris avalanches in the Campania Region: the role of morphological slope discontinuities and the development of failure. Geomorphology 66:237–254
Guerricchio A, Doglioni A, Fortunato G, Galeandro A, Guglielmo EA, Versace P, Simeone V (2012) Landslide hazard connected to deep seated gravitational slope deformations and prolonged rainfall: Maierato landslide case history. Società Geologica Italiana 21:574–576, Roma

Guerriero L, Revellino P, Coe JA, Focareta M, Grelle G, Albanese V, Corazza A, Guadagno FM (2013) Multi-temporal Maps of the Montaguto Earth Flow in Southern Italy from 1954 to 2010. J Maps 9(1):135–145
Guthrie RH, Evans SG (2004) Magnitude and frequency of landslides triggered by a storm event, Loughborough Inlet, British Columbia. Nat Hazards Earth Syst Sci 4:475–483
Haug MD, Sauer EK, Fredlund DG (1977) Retrogressive slope failures at Beaver Creek, south of Saskatoon, Saskatchewan. Canadian Geotech J 14:288–301
Heim A (1932) Landslides and human lives (Bergsturz and Menschenleben). In: Skermer, N. (ed) Bi-Tech Publishers, Vancouver, BC, 196p
Hendron AJ, Patton FD (1985) The Vaiont Slide, a geotechnical analysis based on how geologic observations of the failure surface. U.S. Army Corps of Engineers. Technical Report 85, Number 5, 104p
Highland LM, Bobrowsky P (2008) The landslide handbook: a guide to understanding landslides; U.S. Geological Survey, Circular 1325, 129p
Hoek E, Bray J (1981) Rock slope engineering, 3rd edn. Inst. Mining and Metallurgy, London
Hübl J, Suda J, Proske D, Kaitna R, Scheidl C (2009) Debris flow impact estimation. Proceedings, International Symposium on Water Management and Hydraulic Engineering, Ohrid/Macedonia, Paper: A5
Hungr O (1981) Dynamics of rock avalanches and other types of slope movements. Ph.D. thesis, University of Alberta, Edmonton, 500p
Hungr O (2000) Analysis of debris flow surges using the theory of uniformly progressive flow. Earth Surface Processes and Land-forms 25:1–13
Hungr O, Evans SG, Bovis M, Hutchinson JN (2001) Review of the classification of landslides of the flow type. Environ Eng Geosci VII:221–238
Hungr, O., Dawson, R., Kent, A., Campbell, D. and Morgenstern, N.R. (2002) Rapid flow slides of coal mine waste in British Columbia, Canada. In: Catastrophic Landslides Geological Society of America Reviews in Engineering Geology 15, pp. 191–208
Hungr O, Evans SG (2004a) The occurrence and classification of massive rock slope failure. Felsbau, Vienna, Austria 22:16–23
Hungr O, Evans SG (2004b) Entrainment of debris in rock avalanches; an analysis of a long run-out mechanism. Bulletin, Geological Society of America, no. 9/10, 116:1240–1252
Hungr O, McDougall S, Bovis M (2005) Entrainment of material by Debris Flows. In: Jakob M, Hungr O (eds) Debris flow hazards and related phenomena Chapter 7. Springer, Heidelberg, pp. 135–158 (in association with Praxis Publishing Ltd)
Hutchinson JN (1961) A landslide on a thin layer of quick clay at Furre, Central Norway. Géotechnique 11:69–94
Hutchinson JN (1968) Mass movement. In: Fairbridge RW (ed) Encyclopedia of geomorphology. Reinhold Publishers, New York, pp 688–695
Hutchinson JN (1988) General report: morphological and geotechnical parameters of landslides in relation to geology and hydrogeology. In: Proceedings of the 5th International Symposium on Landslides, Lausanne, 1:3–35
Hutchinson JN (1991) Periglacial and slope processes. In: Forster A, Culshaw MG, Cripps JC, Little JA, Moon CF (eds) Quaternary engineering geology (Edinburgh, 1989). Geological Society Engineering Geology Special Publication No. 7, 283–331
Hutchinson JN (1992a) Landslide hazard assessment. In: Bell DH (ed) Proc. 6th Inter. Sym. on Landslides. Christchurch, N.Z. pp. 1805–1842
Hutchinson JN (1992b) Flow slides from natural slopes and waste tips, in Proceedings, 3rd National Symposium on Slopes and Landslides. La Coruna, Spain, pp 827–841
Hutchinson JN (2002) Chalk flows from the coastal cliffs of north-west Europe. In: Catastrophic Landslides, Evans SG, DeGraff JV (eds) Geological Society of America, Reviews in Engineering Geology XV, pp. 257–312
Hutchinson JN, Bhandari RK (1971) Undrained loading, a fundamental mechanism of mudflows and other mass movements. Geotechnique 21:353–358
Hutchinson JN, Prior DB, Stephens N (1974) Potentially dangerous surges in an Antrim mudslide. Q J Eng Geol 7:363–376
Hutchinson JN, Bromhead EN, Lupini JF (1980) Additional observations on the Folkestone Warren landslides. Q J Eng Geol London 13:1–31
ICIMOD (2011) Glacial lakes and glacial lake outburst floods in Nepal. International Centre for Integrated Mountain Development, Kathmandu, 109p
International Geotechnical Society's UNESCO Working Party on World Landslide Inventory (WP/WLI) (1990) A suggested method for reporting a landslide. Bull Inter Assoc Eng Geol 41:5–12
International Geotechnical Society's UNESCO Working Party on World Landslide Inventory (WP/WLI) (1991) A suggested method for a landslide summary. Bull Intern Assoc Eng Geol 43:101–110
International Geotechnical Society's UNESCO Working Party on World Landslide Inventory (WP/WLI) (1993a) A suggested method for describing the activity of a landslide. Bull Intern Assoc Eng Geol 47:53–57
International Geotechnical Society's UNESCO Working Party on World Landslide Inventory (WP/WLI) (1993b) A multi-lingual landslide glossary. Bitech Publishers, Vancouver, 59p
International Geotechnical Society's UNESCO Working Party on World Landslide Inventory (WP/WLI) (1994) A suggested method for describing the causes of a landslide. Bull Intern Assoc Eng Geol 50:71–74
International Geotechnical Society's UNESCO Working Party on World Landslide Inventory (WP/WLI) (1995) A suggested method for describing the rate of movement of a landslide. Bull Inter Assoc Eng Geol 52:75–78
Jakob M (2000) The impacts of logging on landslide activity at Clayoquot Sound, British Columbia. Catena 38:279–300
Jibson RW (2005) Landslide hazards at La Conchita, California. U.S. Geological Survey Open-File Report 2005–1067
Keefer DK, Johnson AM (1983) Earthflows: morphology, mobilization and movement. USGS Professional Paper 1264
Kieffer DS (2003) Rotational instability of hard rock slopes. Felsbau 21:31–38
King JP, Loveday I, Schuster RL (1989) The 1985 Bairaman landslide dam and resulting debris flow Papua New Guinea. Q J Eng Geol Hydrogeol 22:257–270
Koppejan AW, van Wamelen BM, Weinberg LJH (1948) Coastal flow slides in the Dutch Province of Zeeland. Procs., 2nd International Conference on Soil Mechanics and Foundation Engineering, Rotterdam, Holland, 5:89–96
Lacerda WA (2007) Landslide initiation in saprolite and colluvium in southern Brazil: field and laboratory observations. Geomorphology 87:104–119
Larsen MC, Wieczorek GF (2006) Geomorphic effects of large debris flows and flash floods, northern Venezuela, 1999. Z. Geomorph. N.F. suppl.-vol. 145:147–175. Stuttgart, Berlin
Lavigne F, Suwa H (2004) Contrasts between debris flows, hyperconcentrated flows and stream flows at a channel of Mount Semeru, East Java, Indonesia. Geomorphology 61:41–58
Lefebvre G (1995) Collapse mechanisms and design considerations for some partly saturated and saturated soils. In: Derbyshire E. et al. (eds) Genesis and properties of collapsible soils. Kluver Academic Publishers, pp.361–374. Nelson, J.D. and Miller, D.J., 1992.. Expansive Soils. John Wiley, NY
Leroueil S, Locat J, Vaunat J, Picarelli L, Lee H, Faure R (1996) Geotechnical characterization of slope movements. In: Senneset K (ed) Landslides. Balkema, Rotterdam 1:53–74
Leroueil S, Locat A, Eberhardt E, Kovacevic N (2012) Progressive failure in natural and engineered slopes. In: Eberhardt E, Froese C, Turner AK, Leroueil S (eds) Landslides and Engineered Slopes. Proceedings, 11th International Symposium on Landslides, Banff, 1:31, CRC Press, Boca Raton
Locat A, Leroueil S, Bernander S, Demers D, Jostad HP, Ouehb L (2011) Progressive failures in eastern Canadian and Scandinavian sensitive clays. Canadian Geotech J 48:1696–1712
Locat J, Lee HJ (2002) Submarine landslides: advances and challenges. Canadian Geotech J 39:193–212
Londe P (1965) Une méthode d'analyse à trois dimensions de la stabilité d'une rive rocheuse. Annales des Ponts et Chaussées, Paris, pp 37–60
Lutton RJ, Banks DC, Strohm WE (1978) Panama Canal slides. In: Voight B (ed) Rockslides and avalanches, vol. 2. Elsevier, Amsterdam
McKenna GT, Luternauer JL, Kostaschuk RA (1992) Large-scale mass-wasting events on the Fraser River delta front near Sandheads, British Columbia. Canadian Geotech J 29:151–156
McRoberts EC, Morgenstern NR (1974) The stability of thawing slopes. Canadian Geotech J 11:447–469
Matheson DS, Thomson S (1973) Geological implications of valley rebound. Can J Earth Sci 20:961–978
Meyerhof GG (1957) The mechanism of flow slides in cohesive soils. Géotechnique 7:1–9
Mencl V (1966) Mechanics of landslides with non-circular sliding surfaces with special reference to the Vaiont Slide. Gétechnique 16:329–337
Mollard JD, Janes JR (1984) Airphoto interpretation and the Canadian landscape. Energy, Mines and Resources, Toronto, Canada
Montgomery DR, Dietrich WE (1994) A physically based model for the topoographic control on shallow landsliding. Water Resour Res 30:1153–1171
Morgenstern NR (1967) Submarine slumping and the initiation of turbidity currents. In: Marine Geotechnique, Richards AF (ed) University of Illinois Press: 189–220
Morgenstern NR (1992) The evaluation of slope stability: a 25 year perspective. In: Seed RB, Boulanger RW (eds) Stability and performance of slopes and embankments, ASCE Geotechnical Special Publication 31, 1:1–26
Morgenstern NR, Tschalenko JS (1967) Microscopic structures in kaolin subjected to direct shear. Géotechnique 17:309–328

Moser M (1996) The time-dependent behaviour of sagging slopes (talzuschübe). In: Senneset K (ed) Proceedings, 7th. International Symposium on Landslides. Balkema, Rotterdam, 2:809–814
Nemčok A (1982) Zosuvy v Slovenskych Karpatov (Landslides in the Slovak Carpathians). Slovak Academy of Sciences, Bratislava
Nichol S, Hungr O (2002) Brittle and ductile toppling of large rock slopes. Canadian Geotech J 39:1–16
O'Loughlin CL (1972) A preliminary study of landslides in the coast mountains of southwestern British Columbia. In: Slaymaker HO, McPherson HJ (eds) Mountain geomorphology, geomorphological processes in the Canadian Cordillera. B.C. Geographical Ser. 14:101–11. Tantalus Research Limited, Vancouver, BC
Picarelli L, Olivares L, Comegna L, Damiano E (2008) Mechanical aspects of flow-like movements in granular and fine-grained soils. Rock Mech Rock Eng 41(1):179–197
Picarelli L, Russo C (2004) Mechanics of slow active landslides and interaction with man-made works. Landslides. Evaluation & Stabilization, 9th International Symposium on Landslides. In: Lacerda WA, Ehrlich M, Fontoura SAB, Sayao ASF (eds) Rio de Janeiro (28 June –2 July, 2004), 1141–1176. A.A. Balkema, Rotterdam
Picarelli L, Urciuoli G, Ramondini L, Comegna L (2005) Main features of mudslides in tectonized highly fissured clay shales. Landslides 2(1):15–30
Pierson TC (1986) Flow behavior of chanellized debris flows, Mount St. Helens, Washington. In: Abrahams AD (ed) Hillslope processes. Allen and Unwin, Boston, pp 269–296
Pierson TC (2005) Hyperconcentrated flow—transitional process between water flow and debris flow. In: Jakob M, Hungr O (eds) Debris flows and related phenomena, vol 8. Springer, Heidelberg, pp 159–196
Pierson TC, Janda RJ, Thouret J-C, Borrero CA (1990) Perturbation and melting of snow and ice by the 13 November 1985 eruption of Nevado del Ruiz, Colombia, and consequent mobiliztion, flow, and deposition of lahars. J Volcanol Geotherm Res 41:17–66
Plafker G, Ericksen GE (1978) Nevados Huascarán avalanches, Peru. In: Voight B (ed) Rockslides and avalanches. Elsevier, Amsterdam, 1:277–314
Postma G (1986) Classification for sediment gravity-flow deposits based on flow conditions during sedimentation. Geology 14:291–294
Roberts NJ, Evans SG (2013) The gigantic Seymareh (Saidmarreh) rock avalanche, Zagros Fold-Thrust Belt, Iran. J Geol Soc. doi:10.1144/jgs2012-090
Revellino P, Grelle G, Donnarumma A, Guadagno FM (2010) Structurally controlled earth flows of the Benevento province (Southern Italy). Bul Eng Geol Environ 69:487–500
Sassa, K., 1985, The mechanism of debris flows. In: Proceedings, 11th International Conference on Soil Mechanics and Foundation Engineering, San Francisco, 1:1173–1176
Sassa K (1999) Introduction. In: Sassa K (ed) Landslides of the world. Kyoto University Press, 3–18
Sassa K (2000) Mechanism of flows in granular soils. In: Proceedings of the International Conference of Geotechnical and Geological Engineering, GEOENG2000, Melbourne, 1:1671–1702
Saunders I, Young A (1983) Rates of surface processes on slopes, slope retreat and denudation. Earth Surface Processes and Landforms 8:473-501
Savigny KW, Morgenstern NR (1986) Creep behaviour of undisturbed clay permafrost. Canadian Geotech J 23:515–527
Schumm SA, Chorley RJ (1964) The fall of threatening rock. Am J Sci 262:1041–1064
Schuster RL (Editor) (1986) Landslide dams. Geotechnical special publication no. 3, American Society of Civil Engineers, New York, 164p
Schuster RL, Highland LM (2001) Socioeconomic and environmental impacts of landslides in the western hemisphere. U.S. Geological Survey Open-File Report 01–0276
Schuster RL, Salcedo DA, Valenzuela L (2002) Overview of catastrophic landslides of South America in the twentieth century. In: Evans SG, DeGraff JV (eds) Catastrophic Landslides, Geological Society of America, Reviews in Engineering Geology XV, pp.1–34
Seed HB, Wilson SD (1967) The Turnagain heights Landslide, Anchorage, Alaska. ASCE Journal, 95-SM4:325–353
Seed HB, Lee KL, Idriss IM, Makdisi F (1973) Analysis of the slides in the San Fernando Dams during the earthquake of Feb. 9, 1971, Earthquake Engineering Research Center 73–2. University of California, Berkeley
Sharpe CFS (1938) Landslides and related phenomena. Columbia University Press, NY, 1370
Skempton AW, Hutchinson JN (1969) Stability of natural slopes and embankment foundations. In: Proceedings, 7th. International conference of soil mechanics and foundation engineering, Mexico, State of the Art volume, 291–340
Slingerland RL, Voight B (1979) Occurrences, properties, and predictive models of landslide-generated impulse waves, Developments in geotechnical engineering, rockslides and avalanches, Vol. 2, In: Voight B (ed), Elsevier, Amsterdam, pp. 317–397
Stini J (1910) Die Muren. Verlag der Wagner'shen Universitätsbuchhandlung, Innsbruck (Debris flows, English translation by M. Jakob and N. Skermer, 1997, EBA Engineering Consultants, Vancouver, Canada, 106p)
Strouth A, Eberhardt E (2009) Integrated back and forward analysis of rock slope stability and rockslide runout at Afternoon Creek, Washington. Canadian Geotech J 46:1116–1132
Swanston DN (1974) Slope stability problems associated with timber harvesting in mountainous regions of the Southwestern United States, U. S. Department of Agriculture, Forest Service General Technical Report PNW-021. U. S. Department of Agriculture, Washington, DC
Terzaghi K (1950) Mechanics of landslides (Berkey volume). Geological Society of America, New York, pp 83–124
Terzaghi K (1957) Varieties of submarine slope failures, NGI Publication No. 25. Norwegian Geotechnical Institute, Oslo
Terzaghi K, Peck RB (1967) Soil mechanics in engineering practice, 2nd edn. Wiley, New York, 729p
Turner AK, Schuster RL (2013) Rockfall: characterization and control. Transportation Research Board, Washington, D.C., 658p
Vallance JW (2005) Volcanic debris flows. In: Jakob M, Hungr O (eds) Debris flows and related phenomena, vol 10. Springer, Heidelberg, pp 247–271
VanDine DF (1985) Debris flows and debris torrents in the southern Canadian Cordillera. Canadian Geotech J 22:44–68
Varnes DJ (1954) Landslide types and processes. In: Eckel EB (ed) Landslides and engineering practice, special report 28. Highway research board. National Academy of Sciences, Washington, DC, pp. 20–47
Varnes DJ (1978) Slope movement types and processes. In: Schuster RL, Krizek RJ (eds) Landslides, analysis and control, special report 176: Transportation research board, National Academy of Sciences, Washington, DC., pp. 11–33
Varnes DJ, Savage W (eds.) (1996) The Slumgullion earth flow: a large-scale natural laboratory. U.S. Geological Survey Bulletin 2130
Wang B, Paudel B, Li H (2009) Retrogression characteristics of landslides in fine-grained permafrost soils, Mackenzie Valley, Canada. Landslides 6:121–127
Whalley WB (1984) Rockfalls. In: Brunsden D, Prior DB (eds) Slope instability. Wiley, New York, pp. 217–256
Zaruba Q, Mencl V (1969) Landslides and their control. Elsevier, New York, 238p
Zhang D, Wang G (2007) Study of the 1920 Haiyuan earthquake-induced landslides in loess (China). Eng Geol 94:76–88
Zhang ZY, Chen SM, Tao LJ (2002) The sale mountain landslide, Gansu Province, China. In: Evans SG, DeGraff JV (eds) Catastrophic landslides, Geological Society of America, Reviews in Engineering Geology XV, pp. 149–173
Zischinsky U (1969) Uber Sackungen. Rock Mechanics 1:30–52
Zweifel A, Zuccalà D, Gatti D (2007) Comparison between computed and experimentally generated impulse waves. J Hydraulic Eng 133(2):208–216

O. Hungr (✉)
University of British Columbia,
Vancouver, Canada
e-mail: ohungr@eos.ubc.ca

S. Leroueil
Université Laval,
Québec, Canada

L. Picarelli
Seconda Università di Napoli,
Caserta, Italy