Earth process and geological hazards pdf

You work off campus? Learn more about our Remote Access Options Volume 57, Number 2 Large Earthquakes initiate surface process chains that last much longer than short moments of strong shaking. The most moderate and large earthquakes trigger landslides, from small soil cover malfunctions to massive and devastating rock avalanches. Some landslides dam rivers and lakes seize, which can collapse days later, and mountain flood valleys for hundreds of kilometers downstream. Landslide deposits on the slopes can be remobilized during heavy rains and can evolve into debris streams. Cracks and fractures can form and widen on mountain ridges and flanks, promoting the increased frequency of landslides that last for decades. The more gradual impact involves the washing of excess debris downstream of rivers, which can generate bank erosion and the accumulation of the flood, as well as channel avulsions affecting flood frequency, settlements, ecosystems and infrastructure. Finally, earthquake sequences and their geomorphic consequences alter mountain landscapes on both the human and geological scales. Two recent events have attracted intense research into earthquake-induced landslides and their consequences: magnitude M 7.6 Chi-Chi, Taiwan earthquake in 1999 and M 7.9 Wenchuan earthquake, China in 2008. Using data and information from these earthquakes and several others, we analyze how such events initiate processes that change mountain landscapes, highlight research gaps, and suggest ways to a more complete understanding of the seismic effects on the Earth's surface. Strong earthquakes in mountainous regions trigger chains of events that alter mountain landscapes over the days, years and millennia. The earthquake can cause many tens of thousands of landslides on the steep slopes of the mountains. Some of these sudden slope failures can block rivers and form temporary lakes that can collapse later and cause huge flooding. Other landslides move more slowly, in some cases in a stop-start mode during heavy rains or earthquake aftershocks. Debris from these landslides can block the canals, and during heavy rains, debris can be transported downstream for many kilometers, with catastrophic consequences. New landslides tend to happen more frequently than usual for months to years following an earthquake, since strong ground shaking has fractured and weakened slopes. Other effects of large earthquakes can last, in various forms, over geological time scales. Over the past two decades, our understanding of these problems has advanced thanks to the detailed study of the chi-chi earthquake in Taiwan in 1999 and the 2008 earthquake in Wenchuan in China. and we discuss the results of research on these earthquakes and other earthquakes and explain what we have learned, what we need to know and where we should be leading future studies. Landslides triggered by the earthquake and their consequences are major hazards in the mountains. Destructive earthquakes in the (M 7.6, 1999), China (M 7.9, 2008), Nepal (M 7.8, 2015) and New Zealand (M 7.8, 2016) caused widespread land slipping and prompted researchers to study with increasing detail how earthquake landslides are eroding landscapes. Simonett's early works (1967) and Pears and Watson (1983, 1986) recognized the importance of earthquakes in the sharp increase in rates of erosion, sediment transport and deposition. Detailed work on the 1999 chi-chi earthquake, Taiwan (e.g. Dadson et al., 2004) proposed that earthquakes can instantly change mountain landscapes by triggering the landslide movement of soils, debris and rocklandslides involving 103-1010 m3; more gradual redness of these free materials in the affected areas continues for a period of time after the trigger earthquake. The magnitude and depth of the earthquake are first-order controls on the degree of disturbance of the landscape, modulated by the character of seismic input waves, topography, rock mass properties, groundwater conditions and other factors (Fan, Scaringi, et al., 2018; Gorum et al., 2011). Seismic agitation is an instantaneous disturbance, but its effects are subsiding, and the cumulative effects of earthquakes can leave a mark in the evolution of mountain landscapes (Hovius et al., 2011; Marc, Hovius, & Meunier, 2016). A cascade of processes actually occurs after a large continental earthquake (Figure 1). These processes bring different degrees of danger, which pose a risk when interacting with the human world. Geological hazard chains triggered by a strong continental earthquake and revised in this paper. Causal relationships between hazards are indicated. Red background shows different types of coseismic landslides; the blue background indicates the post-seismic cascade of hazards in days to years later; and the yellow background represents the long-term impact of an earthquake, years to decades later and perhaps longer. Spatial frequency distributions of earthquake landslides are influenced by several factors, including source materials (e.g. soil/regolit cover or the underlying intact base rock type), movement mechanisms (slip, fall, flow or combinations), soil movement characteristics and dimensions (in terms of the surface of the form plane involved, the depth and volume moved; Figure 1, red background). Shaking can also weaken the slopes throughout the landscape and make them more prone to delayed failure (increased rate of occurrence). A few days after the earthquake (blue background), landslide dams formed by landslides triggered by the earthquake can pierce and generate flooding. Remobilisations and coalescence of landslide debris as a result of heavy rains can generate debris. Years to decades later (yellow background), floods may continue, albeit less frequently. Delayed slope failures may also occur, and slow processes, such as river aggradation, will become apparent as debris generated by the earthquake gradually moves downstream. Description and and around these processes and their links to the effect of the case form the structure of this review, as set out in section 1.1. Many landslide disasters of 1900 have had an earthquake origin (Chen et al., 2012; Froude & Petley, 2018; Gorum et al., 2011; 2.1). Investigating the patterns of these processes is essential for risk mitigation, emergency response, reconstruction and increased resilience (Das et al., 2018; Flentje & Chowdhury, 2018; Petley, 2011). The integration of these processes on a longer time scale reveals how earthquakes form mountain belts and providediagnostics for identifying past earthquakes in sediments and relief forms. A large body of literature on specific aspects of earthquake-induced surface processes has grown over the past five decades. More than a thousand articles have been published presenting case studies, models and discussions on the impact of the 2008 Wenchuan earthquake (Fan, Juang, et al., 2018). However, to date they have not covered the full role of earthquake-induced landslides and their consequences in a variety of landscapes (and seismic models) and in short- and long-term time intervals. Only a small number of specialised analyses and meta-analyses have been published that have addressed specific aspects of this topic. The most recognised works, which were relevant to the preparation of our comprehensive review, are reported in Table 1. Table 1. A summary of key papers, analyses and meta-analyses on earthquake landslides and related topics Key references Summary and analysis of reports on earthquake landslides and Keefer snow and ice avalanches (1984, 1994, 2002); Podolskiy et al. (2010a) Guidelines for preparing the inventories of landslides triggered by the Harp et al. earthquake (2011); Xu (2015) Space models and statistics on the size of landslides triggered by the Malamud et al. earthquake (2004); Marc et al. (2017); Marc, Hovius and Meunier (2016) Physical methods for analyzing the stability of the Jibson seismic slope (2011) Physical and statistical assessments of the hazards of landslides, representing seismic agitation Godt et al. (2008); Jibson et al. (1998, 2000); Nowicki Jessee et al. (2018); Nowicki et al. (2014) Patterns of landslides near and far from the landslides triggered by the Delgado earthquake et al. (2011); Gorum et al. (2011); Meunier et al. (2007, 2008) Post-seismic landslide rates Marc et al. (2015); Fan et al. (2019) Export of sediments from seismically active mountains and mass balance of Hovius et al. earthquakes (1997, 2000); Marc, Hovius and Meunier (2016) Seismic shaking can trigger landslides of several sizes, from small surface soil failures to avalanches of large and deep (Figure 1, red background). We use the expressions coseismic landslides or earthquake-triggered landslides (EQTLs) to refer to them. Their distribution depends on seismic wave patterns, geology and topography, and earthquakes can produce hot spots with high density of landslides. In 2, we explain the current knowledge about EQTLs and how to accurately identify, map and analyse and fully their distribution (Section 2.1); point 2.2 discusses the processes by which seismic agitation initiates slope failure. Section 3 is dedicated to post-seismic processes in rivers loaded with coseismic debris (Figure 1, blue background). Section 3.1 summarizes the characteristics of dams and landslide lakes. Section 3.2 describes how the postseismic landscape is sensitive to storms so that more landslides can occur than normal in the months-years after the earthquake. Weakened and fractured agitation hills may experience accelerated degradation: Weather continues to rise, cracks form or expand and post-seismic landslides are initiated (Figure 1). Section 4 discusses improved instruments for earthquake-induced hazard and risk assessment. The cascade of surface processes initiated by an earthquake has the potential to cause damage, disrupt lifelines and services and cause loss of life; a comprehensive hazard and risk assessment must recognise both the immediate and prolonged consequences of earthquakes. Section 5 refers to the cascade of sediments after an earthquake, from the first years when sediment mobility is at its peak until its eventual decrease in ambient rates. An important related question relates to how long it takes for coseismic debris to evacuate the affected area from the rest stored in place. Sediment deposits can be valuable archives for paleoseismic research. Section 6 widens the image of earthquake-induced landscape erosion, taking into account the volumes and fate of sediments (Figure 1, yellow background) produced over several seismic cycles to estimate their contribution to the geological evolution of mountain landscapes. In Section 7, we summarize our current understanding of how earthquakes change mountain landscapes and highlight some perspectives for future research. We offer four technical supplements that present EQTL inventory analysis techniques (support information section S1), land slip runout modeling (Section S2), mechanisms and scrap flow modelling (Section S3), as well as risk mitigation strategies (Section S4). We also include a glossary of technical terms and acronyms. We focus on terrestrial settings, especially mountain areas. We do not consider the submarine landslide, the generation of tsunamis through coastal or offshore earthquakes and their geomorphological and environmental consequences, or earthquake-volcano interactions. These are extensive research topics (Avouris et al., 2017; Dawson, 1994; Hill et al., 2002; Linde & Sacks, 1998; MacInnes et al., 2009; Manga & Brodsky, 2006; et al., 2011) beyond the scope of our review, as well as studies on tsunamis generated by EQTL in mountain lakes (e.g. Ichinose et al., 2000; Kremer et al., 2012; Schnellmann et al., 2002) and the site effects of seismic agitation due to soil and rock types (Bhattacharya (Bhattacharya al., 2011; Evans et al., 2009; Huang & Yu, 2013; Kayen et al., 2013; Wang et al., 2014; Zhang & Wang, 2007). Coseismic landslides, or EQTLs, are mass movements triggered within seconds of strong ground shaking. These may include small, shallow soil failures and rock falls; large, deep and slide slumps; and fast moving, devastating rock avalanches. All the mass movements of the descending slope that occur afterwards are called post-seismic landslides. The key to understanding the coseismic danger of landslide is both to understand the regional patterns of the entire collection of landslides triggered by a particular earthquake, and to understand the trigger ing and leakage processes at a local or even laboratory scale. In the next section, we will discuss different ways of coseismic landslides are studied and what has been learned so far from these studies, from the regional and working perspective to the laboratory scale. Detailed inventories are essential to study the distribution of coseismic landslides, to assess the main slope failure mechanisms and to generate maps of earthquake-induced landslide hazards. In practice, it can be very difficult to distinguish the coseismic landslides from the mass movements that were generated either before the earthquake or by aftershocks in the days/weeks between the occurrence of the earthquake and the purchase of images used for mapping. The challenge is to generate so-called event-based inventories (Guzzetti et al., 2012) or regional events with landslide events (Crozier, 2005) that describe mass movements triggered by a single earthquake. The generation of EQTL inventories has received a lot of attention in recent decades. Keefer (1984) studied the landslides triggered by 40 major historic earthquakes from the 1811/1812 earthquake in New Madrid, Missouri (three M ~ 7.5 events) to the 1980 earthquake of Mammoth Lakes, California (M 6.1). His study correlated the magnitude of the earthquake with the area where the landslides were triggered, the maximum epicentral distance and the maximum defect distance; some of these data were subsequently updated and refined (Keefer, 2002). Similar efforts were made by Rodrıguez et al. (1999) and national databases were generated (Hancox et al., 2002; Papadopoulos & Plessa, 2000; Prestininzi & Romeo, 2000). EQTL mapping uses at least four methods (Guzzetti et al., 2012; Keefer, 2002; Soeters & van Westen, 1996; van Westen et al., 2008; Xu, 2015): (1) visual image interpretation, (2) (semi)automated classification based on spectral characteristics, (3) (semi)automated classification based on elevation differences and (4) field investigation (Table 2). Table 2. general techniques for creating earthquake-induced landslide inventories (After van Westen et al.,) Group Advantages Specific Technical Disadvantages Advantages Visual image interpretation Expert interpretation allows detailed and accurate mapping of landslides as polygons, with such as the type of landslide and the differentiation between the parts of erosion, transport and accumulation. More new developments can accelerate mapping, such as collaborative web mapping, the use of social networks, UAV images, and the Google Earth history viewer. The subjective method may give different results depending on the experience and capacity of the mapper. It is time-consuming and inventories only become available after a few weeks/months. Spatial accuracy can be problematic if mappers use georeferenced images incorrectly. There may be large differences in mapping between different mapping teams for the same area. Video UAV or images Detailed mapping of landslides in small areas. They might be available as overshoot videos, individual images, or 3-D views of point clouds and ortho-rectified images. Apply only on small areas. Permission to use UAVs can be problematic in many areas. Clouds, large differences in altitude and wind can influence the investigation. Aerial video or images from helicopter/airplane Cover larger areas and flight height can be adjusted to get more detailed images. Video coverage is also possible during the same flight. Stereoscopic images are ideal for mapping the inventory of landslides. Very expensive to rent helicopter/plane. Cloud cover can be a major obstacle. Use the history viewer in Google Earth Pro or collaborative web mapping. Direct comparison of images from different periods. There are no additional costs for purchasing images. Specific tools for collaborative web mapping of landslides are now available, such as NASA Landslide Reporter. Geometric distortion can be a problem, and converting from Google Earth KMZ to GIS is cumbersome. Post-event images may not be available or may have cloud problems. High-resolution satellite images cover large areas and can be purchased at different times to show the evolution of landslides. They can be converted into stereoscopic images for optimal interpretation. Persistent cloud coverage may prevent the image from being purchased. High costs involved. High-resolution shaded relief maps (e.g. from LiDAR or UAV) Shaded relief maps from empty surface models derived from LiDAR are best for mapping landslide shapes even under forest cover. Stereo-image interpretation of these images is the ideal tool for detailed interpretation of landslides. Obtaining LiDAR data on large areas in a postseismic situation is very expensive and processing takes time. Classification (semi)automated based on spectral characteristics Rapid assessment of areas affected by regions and usually the best approach for a quick assessment shortly after the earthquake. Some advanced methods also allow the identification of types of landslides. Quick developments using machine learning and cloud computing allow for better results. Requires cloudless images, which can be problematic to acquire in certain areas and seasons. Satellite images can be expensive, although lower costs and lower-resolution alternatives are free. Wrong classification of areas, areas, landslide areas. Difficult to separate areas of amalgamated landslide. Very limited information on detecting changes Relatively rapid method of analysis of changes in land cover using pre- and post-seismic images and analysing changes in the normalised green vegetation index for multispectral images or grayscale in panchromatic images (black and white). Requires high-quality and high-resolution spectral images before and after the earthquake, which can be problematic due to cloud cover. Pixel-based methods Quick method of detecting empty areas after an earthquake, based on normalized green vegetation index. Machine learning and cloud computing approaches offer rapid classification opportunities in the future. Cloud coverage during data acquisition. Excessive simplification of landslide areas. Confusion with other empty land use areas. Analyze object-based images Using spectral information in combination with other features (for example, shape and slope) to outline individual polygons with similar layouts and remove false positives based on other features. Types of landslides can also be identified. The scale parameter that determines the size of individual polygons is difficult to determine. Wrong classification is common. (Semi)automated classification based on elevation differences Comparison of digital high-resolution elevation models allows the calculation of small elevation changes caused by the depletion or accumulation of landslides. It can also be used as a monitoring tool in post-seismic situations. Pre-earthquake digital elevation models are generally of low quality, making it difficult to quantify changes. High-resolution post-seismic DEMS are costly and time-consuming to generate on large areas using LiDAR. Motion Structure Photogrammetric technique for generating points clouds and digital elevation patterns from images derived from UAVs. Quick and fast method. UAV images can only be generated for relatively small areas. In some areas, it is difficult to obtain permits for flying UAVs. Photogrammetry Photogrammetric techniques for generating digital elevation models from overlapping aerial photographs or satellite images. Cover large areas with good accuracy. Image purchase costs can be high, and cloud-free images can be difficult to purchase. Specific stereo satellite images (e.g. Pleiade) are expensive. Laser scanning Earth laser scanning provides clouds of fast points of slopes. Airborne Laser Scan offers very detailed point clouds over the surface, also under the cover of the forest. Obtaining LiDAR data in very large areas in a post-seismic situation is very processing takes a long time. The Interferometric Synthetic Diaphragm Interferometric Interferometric Interferometric Radar provides information about slow-moving landslides on large areas using relatively cheap images. PSInSAR (Permanent Dispersor Interferometry) and other techniques ensure millimeter travel accuracy and is appropriate for post-event monitoring of travel. It cannot be used for rapid landslide events and is therefore not suitable for mapping landslides induced by rapid earthquake. PSInSAR (Permanent Scatters Interferometry) requires sufficient permanent dispersors (buildings and rocks). Field investigation methods Field methods are essential for obtaining evidence of ground slides and form validation of other methods. Field methods allow only the assessment of small areas or even individual landslides. Cannot use to map landslides on large areas. Geomorphological mapping Conventional method of assessing geomorphological and geological decoration of earthquake landslides. Essential for understanding the causal mechanism of landslides. Final validation method for other remote methods. It is only possible to visit small areas. Access to steep areas is difficult or impossible. Usually difficult to get a good overview of phenomena. Mobile GIS Using Mobile GIS and GPS for data collection attribute allows rapid mapping and is used in combination with the above remote methods, from which the inventory of landslides is obtained, which is then validated and adjusted. Device battery life can be problematic, as can the readability of screens in sunny environments. The reception of the GPS signal could be problematic on steep terrain. Data may be lost when the equipment is interrupted. Road studies Relatively fast method of assessing landslide events along road networks. They can also use dashboard cameras and software (for example, Maphillary or Google Street View) to collect data and map later at the office. Only the limitation to roads and mapping of landslides away from roads is limited. Therefore, the inventory will be biased and difficult to use in area-level studies. Roads could be blocked. Interviews Using questionnaires, workshops, and so on, the local population can provide very useful information on how landslides behaved during the earthquake. Essential for understanding the mechanism and impact of the event. The local population could be in shock as a result of the event and may have moved into shelters. They might exaggerate/underplay information for specific reasons. Cross-checking is required. Geophysical studies Wide range of tools available for measuring geometric, geotechnical, geophysical and hydrological properties that are essential for analysing the stability of earthquake landslides. Point-by-point investigations shall provide only on-the-spot investigation information. Difficulty in sites with equipment. Expensive and time-consuming. Note. GIS = Geographical information system; UAV = Unmanned aerial vehicle. Visual image interpretation is the most common approach to generating EQTL inventories (Schmitt et al., 2017; Tanyas, Allstadt, & van Westen, 2018). The interpretation of aerial photographs aided by stereoscopic vision allows the tracking of features and remains a reference point in the mapping of landslides (Guzzetti et al., 2012). However, air-overhead areas affected by earthquakes are expensive, which reduces the completeness of some inventories (Dai et al., 2011; Xu, Xu, & Shyu, 2015). Since the launch of NASA (National Aeronautics and Space Administration) Landsat-1 in 1972, satellite images have provided more systematic coverage of landslides. The stereoscopic capacity and higher spatial resolution (10-20 m) of SPOT images (released in 1986) proved very useful in land landslide mapping, and subsequently LandsatTM and ETM+, IRS 1C/1D and SPOT images were widely used for inventory preparation (Crowley et al., 2003; Gupta & Sava, 2001; Salzmann et al., 2004; Zhou et al., 2002). The entry of Space Imaging Inc. (now DigitalGlobe) into the iKONOS-2 satellite market in 1999 was a milestone in land landslide mapping. With a resolution of 1-m, the image quality rivalled that of aerial photos of 1:10,000 scales (Nichol & Wong, 2005). More recent platforms include commercial satellites such as WorldView (0.31 m), GeoEye (0.41 m), Pleiade (0.5 m) and QuickBird (0.82 m). Today, Planet Constellation offers image resolutions from 0.72 (SkySat) to 3 m (Dove, RapidEye) for a specific daily area. The Copernicus Sentinel-2 mission offers free multispectral images with a resolution of 10 m covering the Earth between 56°S and 84°N, on average every 3 days (Figure 2). Timeline of the development of satellite optical sensors that are used in resource and hazard inventories. Aerial recognition from fixed-wing aircraft or helicopters is a quick, al. selective, way to assess the scope of landliding in an earthquake-affected area (Rosser et al., 2014; Van Dissen et al., 2013), but is not suitable for systematic investigations. Unmanned aerial vehicles (UAVs) can provide detailed videos, images, 3-D-point clouds and ortho-rectified images for parts of the affected area or for individual landslides (C. Tang et al., 2016), but are too limited in range to map large landslides. Google Earth Pro stores historical images that help detect changes and can be used to map EQTL without cloud coverage (Gorum et al., 2013). At present, specific tools are becoming available for mapping landslides, would be NASA Landslide Reporter 2019 (Kirschbaum & Stanley, 2018), although much more needs to be done to make them suitable for collaborative web mapping of EQTL inventories due to lack of recent images and specific tasks per mapper, as is the case in HumanitarianStre OpenEtMap 2019 (Winsemius et al., 2019). The interpretation of satellite or aerial images should ideally involve stereo images and experts trained in identifying, classifying and validating landslides based on image tone, texture, size and model (Soeters & van Westen, 1996). Landslides should be mapped as polygons rather than dots to document their areas and volumes. Where possible, landslide polygons should separate the initiation areas from the drainage areas, as only the source areas are used in the statistical risk of landslide Identifying individual landslides is important, but it is difficult to identify where they merge. Landslides must be classified by type, depth class and river blocking potential (Figure 3). Example of earthquake-induced landslide mapping as polygons with attribute information. The semi-automatic classification of satellite images is a tool that develops rapidly and produces increasingly reliable dimensions of landslides (Fan, Juang, et al., 2018), although careful validation of experts (hence the semi-automatic term) is still required. Image classifications based on spectral image characteristics increase in accuracy through the use of unsupervised and supervised algorithms, machine learning approaches, and object-based image analysis (Anders et al., 2011; Moosavi et al., 2014; Stumpf & Kerle, 2011). Individual landslide polygons can now be successfully outlined using Object-based Image Analysis (K. -T. Chang, Hwang, et al., 2011), although problems remain with the separation of merged landslide deposits. Spectral, topographical and shape measurements may characterize EQTLs according to their type (Martha et al., 2010). Semi-automatic methods also include comparisons of digital elevation models (DEM) (mainly derived from LiDAR- or UAV) and their differences in pixels before and after an earthquake. This requires the removal of artificial objects and vegetation to generate surface patterns. Several nonoptic sensors overcome the cloud cover problem (Williams et al., 2017) while tracking surface deformations at millimeter precision via synthetic aperture radar (Casagli et al., 2016; Guzzetti et al., 2012). Although this method is not suitable for detecting fast, long-running EQTLs, it has the potential to monitor the activity of coseismic landslides in the months and years after an earthquake. The field verification of landslide inventories remains essential for the validation of these remote-detection techniques (Table 2; Harp et al., 2016; Harp & Jibson, 1995, 1996). Although these methods are generally not used as the main tool for mapping the EQTL inventory due to difficulties accessing the entire affected area after an earthquake, they are essential for validating the other methods. Mapping landslides in the field provides detailed data on the type, internal composition and mechanism of failure of landslides, which are usually difficult to assess by tele-detection. Field mapping of landslides is time-consuming and intractable for the thousands of landslides triggered during earthquakes. Despite, or perhaps even due to, the variety of methods, EQTL mapping inventory led to very different results. For example, Parker et al. (2017) used an automatic and visually verified mapping method for landslides triggered by the 2008 Wenchuan earthquake. Their results differ significantly from those based solely on visual interpretation (Fan, Juang, et al., 2018; Gorum et al., 2011; Xu, Xu, Shen, et al., 2014), mainly because adjacent landslides were generally mapped as a single which overestimated the size of landslides and underestimated the number of landslides (Figure 4). Poorly mapped ground sliding locations can also affect the accuracy of spatial queries in terms of causal factors, such as rock or soil type. Comparison of the mapping of landslides triggered by the earthquake in the same area by different groups. Example from the epicentral area of the Wenchuan earthquake of 2008. (a) A post-seismic satellite image showing the landslides that were subsequently triggered and mapped. (b)–(d) presents a point inventory mapped by Gorum et al. (2013) and polygonal inventories (red) mapped by three different groups: (b) Z. Z. Li, Jiao, et al. (2014); (c) Xu, Xu, Yao and Dai (2014); and (d) Fan, Domènech, et al. (2018). In view of the large differences between eqtL inventories, standard criteria for assessing the quality of landslide inventories are desirable (Marc & Hovius, 2015; Tanyas et al., 2017; Xu, 2015). More studies (Guzzetti et al., 2012; Harp et al., 2011; Xu, 2015) the proposed criteria, e.g.(1) all identifiable landslides should be mapped; (2) the boundaries of landslides should be mapped as polygons rather than just as mapping of a location of landslide points; (3) Regional mapping of landslides should cover the entire affected area and the outline of the mapped area should be indicated; (4) Complex landslides should be divided into distinct groups; (5) landslides should be mapped as soon as possible after the earthquake; (6) Stereo coverage or draping over a DEM is preferred to allow oblique views; (7) landslides must be plotted in a designed coordinate system; and (8) visual interpretation is preferable to automatic extraction. Few mapping studies have met all these criteria (e.g. Gorum et al., 2014; Xu et al., 2013; Xu, Xu, & Shyu, 2015), which could explain some of the large differences in size and positional accuracy between inventories (Marc & Hovius, 2015; Xu, Xu, Yao, & Dai, 2014; Figure 4). Many studies have summarized the properties of EQTL inventories (Gorum et al., 2014; Harp et al., 2011; Havenith et al., 2016; Keefer, 1984, 2002; Rodrıguez et al., 1999; Tanyas et al., 2017; Xu, 2015). The most complete is that of Tanyas et al. (2017), which compiled data on 363 earthquakes and gathered 66 digital inventories of coseismic landslides, which were classified as (1) comprehensive digital inventories with full coverage of all landslides and their types; (2) non-comprehensive digital inventories with partial coverage, type descriptions or point data; (3) reports on EQTLs without a digital inventory; (4) reports on EQTLs without any and (5) earthquakes with potential, although no landslides have been reported. Tanyas et al. (2017) also presented associated metadata for digitally available EQTL inventories. The metadata provided contains valuable information for an overall inventory assessment. In view of the metadata, users can decide whether an inventory is appropriate for the purpose of their application. This also warns users of some data limitations. For For in this database, some inventories include landslides triggered by a sequence of earthquakes rather than a single mainshock (e.g. 1980 Mount Diablo and 1993 Finisterre Mountains). Schmitt et al. (2017) has created an open repository to house the digital inventories they are allowed to share through the U.S. Geological Survey ScienceBase platform (Table S1 and Figure 5). The new inventories of landslides are those of 2016 M 7.8 Kaikōura, New Zealand (Massey et al., 2018; Sotiris et al., 2016), 2017 M 6.5 Jiuzhaigou, China (Fan, Scaringi, et al., 2018; Xu, Wang, et al., 2018) and 2012 M 5.7 Yiliang, China (JJ. Zhang, Wang, Zhang, et al., 2014) earthquakes. The largest and most detailed of the inventories is that of the 2008 earthquake in Wenchuan, China, which now contains about 200,000 individual coseismic landslides covering a total area of 1,160 km2 (Xu, Xu, Yao, & Dai, 2014); a sub-invention lists 828 coseismic landslide dams (Fan, van Westen, et al., 2012). Distribution of earthquakes that have inventories of earthquake-generated landslides, digitally available, listed in Table S1 (Additional information, adapted from Tanyas et al., 2017; Schmitt et al., 2017). The data store can be accessed online (. The background is from Amante and Eakins (2009; . As can be seen from Figure 5, EQTL events occurred in most active tectonic mountain areas in the world, with a predominance in warm, temperate and humid regions. However, earthquake-triggered snow and rock avalanches have also occurred in cold regions, where they can travel long distances through ice and snow cover. Glazovskaya et al. (1992) established that 6.2% of the Earth's surface that has a snow depth of more than 30-50 cm, a slope steepness >17°, and a slope height of 20-30 m are prone to snow avalanches. Until recently, very little systematic documentation of earthquake avalanches was made (Chernouss et al., 2006; Podolskiy et al., 2010a). The most comprehensive and recent publication of Podolskiy et al. (2010a) reports 22 historical cases of earthquake avalanches that occurred between 1899 and 2010 in cold Arctic regions and very high mountain areas (e.g. the Himalayas). Important cases include rock and ice avalanches in the Chugach Mountains, triggered by the 1964 Alaska earthquake (Post, 1967; Tuthill & Laird, 1966); the 1,580 large landslides triggered by the 2002 Denali earthquake over alaska's glaciated terrain (Gorum et al., 2014; Jibson et al., 2004); the 1,448 landslides induced by the melting of snow following the 2004 Mid-Niigata prefecture earthquake (Akiyama et al., 2006) and long-term snow-triggered land during the 2011 Nagano prefecture earthquake in Japan (Has et al., 2012; Yamasaki et al., 2014). Landslide size statistics provide useful contributions for hazard assessment (e.g. Guzzetti et al., 2005) 2005) studies on landscape evolution (e.g. Korup et al., 2012), as probabilities of landslide size, total area and volume can be estimated from their empirical distributions. These size distributions are mostly reported as normalized histograms. Many studies claim that in most EQTL inventories, distributions of medium and large landslide areas follow a reverse power law: (1)where f is frequency density, c is a dimensional constant, A is the area of landslide, and β is a scaling exponent (Figure 6; e.g. Malamud et al., 2004). This relationship was also reported for landslide triggers, such as storms and rapid melting of snow (Malamud et al., 2004), and its apparently universal applicability motivated the application of these statistics to hazard assessments (Guzzetti et al., 2005). Some examples of distributions of probability zones derived from digital inventories of earthquake-triggered landslides (Table S1) mapped for different earthquakes showing (a) the right curves of a double pareto distribution using Rossi et al. code (2012) and (b) the power law fits. Probability and density distributions (grey lines) are plotted for 32 earthquake landslide events in Tanyas, van Westen, et al. (2018). Characteristic characteristics are labelled in inset plots. Some examples of specific earthquake inventories are highlighted in colors to show the range of typical curve shapes (left) and the upper and lower boundaries observed for the exponents of the power law (right). In most EQTL inventories, small landslides depart from the law of power and form a reversal (Figure 6a) under which they are less common (e.g. Malamud et al., 2004; Tanyas, van Westen, et al., 2018). The approximate location where the frequency distribution differs from electrical power legislation is commonly referred to as interruption and is interpreted either as a statistical sampling artifact or as a function of the bulk ownership of the hills; the scaling exponent describes the relevance of larger landslides (Figure 6). Frequency zone distributions can be used to identify a scale of the magnitude of landslide events (Malamud et al., 2004) to characterize the geomorphic severity of an earthquake. Keefer (1984) pioneered this idea considering the total number of landslides turned into a log triggered by an earthquake: An event that triggers landslides 102-103 is classified as 2; 103–104 landslides is classified as a 3, and so on. Malamud et al. (2004) suggested that the completeness of landslide inventories can be assessed by comparing the frequency of a partial inventory with the proposed empirical curves (Section S1). However, using 45 earthquake-induced landslide inventories, Tanyas, van Westen, et al. (2018) found that the empirical curves proposed by Malamud et al. (2004) did not fit many inventories and that the installation was subjective and prone to uncertainty. They updated the method that allowed for more flexible curve shapes with different β values, and proposed a more reproducible and robust way to the extent of the landslide. Reliable magnitudes of landslide events could allow direct comparison of different events that trigger landslides. Tanyas et al. (2019) proposed an equation of regression of the morphometric and seismic variables available globally for the near-real-time prediction of the magnitude of the EQTL event, the total landslide area (Malamud et al., 2004; Tanyas, van Westen, et al., 2018) and volume (Malamud et al., 2004). Compared to the area of landslide, determining the volume of landslides is difficult, as it requires knowledge of the geometry of the pre and post-landslide surface or the underground geometry of the slope failure (Guzzetti et al., 2009). Even techniques such as the decrease of high-resolution DEMs before and after landslides can estimate volumetric variation only because part of the source area may overlap with deposits (C. Tang et al., 2019). Therefore, empirical relationships are commonly used to estimate the volume of landslides. Several authors argue a rightful power relationship between the area affected by landslide (A) and volume (V), i.e.(2) and that the geometry of the landslide is regardless of the specific mechanical properties (Guzzetti et al., 2009; Klar et al., 2011). For example, Guzzetti et al. (2009) examined 677 slides and estimated a scaling exponent (γ) of 1.45. Klar et al. (2011) used the boundary balance principles to explore the relationship between the volume of landslide and the surface and claimed that γ rangebetween 1.32 and 1.38. Laren et al. (2010) observed considerable variations in the scaling exponent with landslide material: the values γ for the soil are 1,1–1,3, and for the base rock are 1,3–1,6. They also showed that the total volume of landslides is very sensitive to γ so that small differences in γ can cause a prediction error of one or more orders of magnitude in the total volume of landslides. Korup et al. (2012) stated that numerical errors in the parameters of the land landslide distribution model can cause uncertainty greater than a factor of 2 in total landslide volumes. They conclude that such minor errors can cause great uncertainty in what or mobilization rates deducted from size statistics. To address this problem, frequency-sized distribution analyses should be performed using reliable EQTL inventories and uncertainties caused by inventory data should be examined in detail. The reliability of landslide size statistics depends on the reliability of the inventory. Four major problems can compromise the quality of the landslide inventory and its size statistics: (1) the accuracy of limit mapping of land, (2) coalescing landslides, (3) overlap of landslide sources and deposits and (4) geometrical distortions in the image and due to topography. Propergeoreferenced images with an appropriate spatial resolution are required for the precise delimitation of landslide boundaries and, therefore, for the accurate assessment of landslide areas. Mappers' level of expertise and time invested in mapping are important, even when an inventory is created using semi-automatic semi-automatic Techniques. Landslide contours can be difficult to map if images are weak or contrast low, or if two or more landslides join together or do so gradually after an earthquake; if there are many such cases, this may also change the distributions of the size of landslides (e.g. Marc & Hovius, 2015; Tanyas, Allstadt, & van Westen, 2018) and volumetric estimates (Z. Li, Jiao, et al., 2014). The factors that control the leakage of landslides differ from those that control the slope failure that creates the area of the source of the landslides (Frattini & Crosta, 2013). Thus, the landslide source area should be treated separately in the analysis of the frequency of landslides, as the landslide deposition side could affect the landslide size statistics, especially in the case of long-distance landslides (Tanyas, van Westen, et al., 2018). However, landslide sources and deposits are rarely mapped separately (e.g. Frattini & Crosta, 2013), especially in EQTL inventories (Tanyas et al., 2017). To address this issue, Tanyas, van Westen, et al. (2018) exclude debris flows from the EQTL inventories examined because of their larger drainage and deposition areas. Fixed ratios between source and warehouse areas (Laren et al., 2010) can be a tough guide to differentiating sources of landslides (e.g. Marc & Hovius, 2015; Marc, Hovius, & Meunier, 2016). Although no clear physical explanation was found (Hergarten, 2003), EQTL inventory distributions are typical of other natural hazards, such as earthquakes and forest fires (e.g. Hergarten, 2003; White et al., 2008). Topography is an explanation of the candidate for the distribution of landslides (e.g. Frattini & Clefta, 2013; Liucci et al., 2017; ten Brink et al., 2009), because topography is the main factor limiting the boundaries of the sliding material. The Valleygussa et al. (2019) argue that the intensity of soil movement has a stronger influence on the distribution of the size of landslides and noted that stronger ground shaking can trigger larger landslides. However, only three out of six eQTL inventories that they examined supported this line argument. The characteristics of soil movement, such as frequency content and duration, could also alter the size distribution (Jibson, 2011). For example, Jibson et al. (2004) reports that the 2002 Denali earthquake triggered significantly lower concentrations of lower rock falls and rock glides compared to earthquakes with magnitudes at least as high. They claimed that this occurred because the earthquake was deficient in high-frequency energy and of the related high amplitude. This hypothesis has not yet been tested and therefore requires further analysis. The exponents of the power law (β values) of landslide inventories range from 1.4 to 3.7 (Stark & Guzzetti, 2009; Tanyas, van Westen, et al., 2018; Van Den Eeckhaut et al., 2007; Figure 7). As regards hazard assessment, a high β indicates a higher proportion of small and medium landslides. On the other hand, a smaller β means a higher proportion of large landslides. Differences in β could regional differences in structural geology, morphology, hydrology and climate (Bennett et al., 2012; Chen, 2009; Densmore et al., 1998; Dussauge-Peisser et al., 2002; Hergarten, 2012; C. Li, Ma, et al., 2011; Sugai et al., 1995); however, a convincing relationship has not yet been found, partly due to uncertainty in the estimate of β, which may be about 40% in some cases (Tayas, van Westen, et al., 2018). Similarly, the relationship between earthquake magnitudes and the exponents of the law of power distributions of landslide sizes remains noisy (Figure 7). The distribution of data between earthquake magnitude and energy value matches density probability distributions, with uncertainties calculated using data from Tanyas, Allstadt and van Westen (2018) for 45 inventories of earthquake landslides. After mentioned above, also the reason for rollover and divergence from the law of power is controversial: Some have claimed that it is controlled by the mechanical properties of the substrate (Bennett et al., 2012; Frattini & Crosta, 2013; Guthrie et al., 2008; Pelletier et al., 1997; Stark & Guzzetti, 2009; Van Den Eeckhaut et al., 2007); others have speculated that it may be in part a mapping artifact caused by a lack of space (Hungr et al., 1999; Stark & Hovius, 2001) or temporal (Tanyas, van Westen, et al., 2018; Williams et al., 2018 resolution. If the rollover is purely a mapping artifact, then refined distribution size models of landslides are required. Tanyas, van Westen, et al. (2018) found that some inventories did not have rollovers (Figure 6a) and noted that both physical explanations and other arguments can work together in forming a rollover. Some studies consider the rollover to be the smallest size of the landslide to which the inventory can be assumed to be complete (Parker et al., 2015; Van Den Eeckhaut et al., 2007). Coseismic landslides are triggered as a result of a complex interaction between several trigger factors, in particular seismicity and predisposing factors, such as soil mass/rock properties, slope geometry and structural characteristics (Table 3). Trigger factors associated with ground shaking are the most important because they control the regional model of EQTL distribution (Nowicki et al., 2014). Table 3. Overview of the triggers and predisposing factors controlling EQTLs with an indication of their relative importance and the feasibility of spatial representation groups/subgroups control factors Importance can be mapped? Examples of references Approaches/data commonly used to represent control factors Trigger factor Seismic effect VH VH Keefer (1984) Seismic acceleration, speed and intensity (Nowicki et al., 2014) Intensity (Jibson, 1993) Depth VH VH Meunier et al. (2007) Duration VH VH Nowicki Jessee et al. (2018) Aftershocks H VH Tiwari et al. (2017) Mecanism focal VH VH Gorum şi Carranza (2015) Ruptură lungime VH VH Marc, Hovius, şi Meunier (2016) Asperitate Asperitate H M Gorum et al. (2014) Surface rupture VH VH C. Xu (2014) Distance from seismic source M VH Kritikos et al. (2015) Topographic amplification VH VL Meunier et al. (2007) Earthquake directionality VH M Roback et al., 2018) Arias intensity H H Jibson (2007) Conditions of predisposition Ground/rock mass properties, slope geometry and structural characteristics Topography Elevation L-M VH Gorum et al. (2011) SRTM, ASTER, ALOS DEM Internal Relief M-H VH Tanyas et al. (2017) Slope Abruptity VH VH Zevenbergen and Thorne (1987) Slope Direction M-H VH Lombardo et al. (2018) Planar and curvature of profile M-H VH Heerdegen and Beran (19 82) Relative position of slope M-H VH Böhner and Selige (2006) Moisture Topographic Index M VH K. J. Beven and Kirkby (1979) Lithological Geology Types VH VH Nadim et al. (2006) Global Lithology Map (Hartmann & Moosdorf , 2012) Characteristics of rocks (geotechnical) VH L-M Duncan et al. (2014) Relationship gaps with topography VH L Ghosh et al. (2010) Discontents and defects VH M-H Chigira and Yagi (2006) Soil types H H Khazai and Sitar (2004) SoilGrids (Hengl et al., 2014) Soil thickness VH M C. Tang et al. (2011) Soil characteristics (geotechnical/hydrological) VH L-M Wyllie and Mah (2014) Hydrology Climate Zone VH VH Tanyas et al. (2017) Bioclimatic regions (Metzger et al., 2013), IMERG (Huffman et al., 2010) and TMMPA (Huffman et al., 2010; Huffman & Bolvin, 2013) Precipitation before the earthquake VH H. B. Wang et al. (2007) Precipitation immediately after the earthquake VH H Sassa et al. (2007) Distance to rivers H VH Tanyas et al. (2019) Land use Land cover/land use M-H VH Yi et al. (2019) GlobCover (Arino et al., 2008) Normalized Difference Vegetation Index M-H VH Song et al. (2012) Land cover changes H H Alcántara-Ayala et al. (2006) Distance to roads M VH Stanley and Kirschbaum (2017) Note. VL = very low; L = low; M = moderate; H = large; VH = very high; SRTM = Shuttle radar topography mission; ASTER = Space-based advanced thermal emission and reflection radiometer; DEM = digital elevation model IMERG = Integrated multi-satellite retrieval for GPM; TMPA = TRMM multi-satellite precipitation analysis. Examples of research investigated the role of these factors in the occurrence of spatially triggered earthquake landslides are also presented. These control factors are highly interconnected. For example, the properties of the soil/rock mass of a slope are mainly controlled by the physical properties of a lithological unit forming the slope. However, in order to assess the mass properties of a slope, other factors, including weather processes, the effects of previous earthquakes/landslides, anthropogenic influence, land cover and land use types (Table 3) should be taken into account. Some of the control factors have Interactions. For example, weather processes affect both the mass of the soil/rock and the geometry of a slope. Another example is that climate and weather interact, because the moisture content can enhance chemical weather. Each the event has unique combinations of these control factors that control the characteristics of each event (Tanyas, van Westen, et al., 2018), such as the locations, types, dimensions and total areas affected, as well as the total density and number of EQTL. For example, the total area of landslides triggered by the 2008 Wenchuan earthquake (M 7.9) was approximately ten times that of the EQTLs triggered by the 2015 Gorkha earthquake (M 7.8), even though they had similar magnitudes and occurred in similar geomorphic and seismothectonic environments. The difference was caused by specific control factors related to the characteristics of the defect rupture and its orientation (Kargel et al., 2016). More efforts are needed to better understand the control factors and their interactions in order to predict the distribution of coseismic landslides in the future. This will require that the control factors be spatially represented so that they can be used in a predictive model. Table 3 provides a general indication of whether the different control factors can be mapped. Assessing slope stability is a geotechnical problem and requires the identification of each component of the driving and resistance forces along a sliding surface. This requires a comprehensive investigation ranging from the physical properties of the slope material to the geometry of the sliding surface, which can be either a discrete surface or a deformation area. The full summaries of these investigations are presented by Wyllie and Mah (2014) and Duncan et al. (2014), among others. In regional studies, in most cases detailed geotechnical data necessary for the deterministic assessment of slope stability are not available. Therefore, researchers prefer either simplified approaches to physical modeling (Jibson et al., 1998, 2000) or statistical methods (Nowicki et al., 2014; Nowicki Jessee et al., 2018) considering proxies for causal factors in regional studies. While the control factors for EQTL are to some extent comparable to those for landslides triggered by precipitation, common factors related to land use, hydrology, soils and geology are not described in detail here. Descriptions of their importance can be found in van Westen et al. (2008) and Reichenbach et al. (2018). The characteristics of the earthquake, such as magnitude and depth, play an important role in the distribution of EQTL (Keefer, 1984; Rodrıguez et al., 1999). However, earthquakes with similar magnitudes may cause different levels of agitation and deterioration (Wald et al., 2003) due to the interaction of other factors, such as the energy released by failure, directionality, topographic amplification and the frequency and duration of the shaking. For example, for earthquakes that have a unilateral rupture of the fault (e.g. earthquakes in and Gorkha of 2015; Roback et al., 2018; Xu, Xu, Yao, & Dai, 2014; Xu et al., 2016), ground movements may be substantially different along the defect. This frequently leads to a higher concentration of coseismic coseismic coseismic along the broken defect. Another factor controlling soil movement is the defect rupture mechanism. Different types of seismogenic defects typically generate different EQTL models (Gorum & Carranza, 2015; Meunier et al., 2013; Father & Grasso, 2013). Figure 8 provides a schematic illustration of the importance of the fault rupture mechanism on the EQTL distribution. Schematic representation of the effect of the defect type on the distribution of landslides triggered by the earthquake. In the case of inverse defects, the density of EQTLs is higher on the suspended wall (Sato et al., 2007; Xu, Xu, Shen, et al., 2014; Xu & Xu, 2012; Figure 8). In the event of a parallel failure on the suspended wall of an inverse seismogenic failure, the area between the two malfunctions usually has a high concentration of EQTLs (Basharat et al., 2016; Kamp et al., 2008; Wang et al., 2002; Xu, Xu, & Shyu, 2015; Xu, Xu, Yao, & Dai, 2014). Landslides triggered by the strike sliding fault tend to have a symmetrical distribution around the fault (Gorum et al., 2014; Xu & Xu, 2014), but in a narrower area (Figure 8). Landslides triggered by a normal malfunction are also more likely to occur on the suspended wall (Xu, Ma, et al., 2018), but there are too few cases to draw a conclusion. The main reasons for eqtl distribution differences due to the difference in failure style are as follows: (1) The highest ground movements tend to occur closer to the surface fault rupture (Abrahamson & Somerville, 1996), and a diving fault surface (reverse or normal) has larger surfaces closer to the surface of the ground than a vertical fault surface (more typical of the strike slide) , resulting in a higher concentration of EQTL on hanging walls (Tatard & Grasso). , 2013); (2) hanging walls can also capture energy, resulting in stronger ground movements due to constructive interference of seismic waves (B. Shi et al., 1998); and (3) the suspended wall wedge is much smaller than the leg wall, so the same force applied to both sides of the defect will lead to higher accelerations in the suspended wall (Oglesby et al., 2000). In addition, although both normal and reverse failure events experience this suspended wall effect, inverse defects tend to generate stronger shaking than normal defects (Oglesby et al., 2000). In addition to the type of failure, Gorum et al. (2014) argued that areas with high density of landslides could be related to asperities that can be observed along the failure plan, which are areas where a large amount of energy is released during breakage (Ruiz et al., 2011) due to maximum friction (Hall-Wallace , 1998). Huang and Fan (2013) took into account the characteristics of the event breaking to understand the distribution of EQTLs for the 2008 Wenchuan earthquake and suggests a control of locking fault junctions, which rarely fail intersections of fault segments (Shen et al., 2009). Huang and Fan (2013) associated areas with high density landslide areas with fault junction areas, where a large amount of energy was released by In order to capture the effect of seismicity taking into account all the control factors listed above, the most representative models can be obtained by numerical simulations (e.g. Imperators & May, 2015; Lee et al., 2009, 2008). However, such approaches have high calculation costs. Given this, the most appropriate data available worldwide for most earthquakes are from the American Geological Survey ShakeMap database (Worden & Wald, 2016), which provides deterministic estimates of ground motion parameters, including peak ground acceleration (PGA), peak ground speed (PGV), modified mercalli intensity and spectral acceleration. Several studies have shown that the probabilities of landslides change substantially when PGA, PGV and Modified Mercalli Intensity are applied as the main seismic factors in the landslide model, as they directly reflect the properties of soil movement (Meunier et al., 2007; Nowicki et al., 2014; Nowicki Jessee et al., 2018). Although seismicity factors are most important in controlling the distribution of EQTLs, predisposing conditions and their interaction with seismicity also play an important role in landliding spatial models. For example, the topographical amplification of the site strongly affects soil movements and therefore the occurrence of coseismic landslides (Meunier et al., 2007, 2008). The grouping of landslides near the slopes ridges, where the topographical amplification is greatest (Davis & West, 1973; Rizzitano et al., 2014) is an example that predisposing factors for EQTL are not necessarily similar to those with precipitation onset. L. Shen et al. (2016) showed that the appearance of the slope can affect eqtl primarily as a result of seismic factors, such as the directions of movement of geological blocks, crust stress and the propagation of seismic waves. These differed from the influence of the appearance of the slope on non-seismic landslides, which relate mainly to differences in vegetation coverage, exposure to sunlight, precipitation and soil moisture (Kamp et al., 2008; Yalcin, 2008). The coupled effect of seismicity and precipitation increases the magnitude of landslide trigger events (Faris & Wang, 2014; Sassa et al., 2007). For example, the Hokkaido region of Japan was struck by a 6.6 M earthquake on September 5, 2018, just after Typhoon Jebi occurred on August 27, 2018 (Yamagishi & Yamazaki, 2018), which recorded over 100 mm of rain in the area where the earthquake caused numerous landslides. There may be an additional effect on seismic agitation from previous precipitation, which may have increased pore pressures and altered slope balance by favouring malfunction mechanisms during soil movement. Other predisposing factors would be the properties of the mass slope geometry and structural characteristics also affect the spatial distribution of landslides. Topographic factors can be easily derived using globally available DEMS, such as the NASA Jet Propulsion Laboratory; Abrams & Hook, 2013; Farr et al., 2007), 2007), Advanced Spaceborne Thermal Emission and Reflection Radiometer Global DEM (Tachikawa et al., 2011), ALOS Global Digital Surface Model (Tadono et al., 2016) or WorldDEM by Airbus 2019. Using these data sources, dems with resolutions up to 12 m can be obtained. On the other hand, the necessary data on other predisposing conditions are generally not available globally in high resolution (e.g. soil type or lithology), although global products become available, such as SoilGrids (Hengl et al., 2014) or OneGeology (Laxton et al., 2010). The association between specific conditions (e.g. land use and soil types) and susceptibility to landslides is not simple to characterize globally (e.g. Stanley & Kirschbaum, 2017). For example, a global geological map is provided by Sayre et al. (2014) with a resolution of 250-m. However, although units may have the same geological description, they may have a completely different role in initiating landslides due to differences in associated structural and geotechnical characteristics. Unfortunately, neither structural characteristics nor detailed lithographical data are available globally at larger mapping scales (greater than 1:100,000). As a result, structurally controlled failures that can create large landslides (Chigira & Yagi, 2006) are not taken into account in multivariate regional analysis. Similarly, hydrological conditions are not generally taken into account as no data on groundwater or high-resolution rainfall are available globally, with some rare but insightful exceptions (e.g. Kogure & Okuda, 2018). As a result, all available statistical models that allow us to understand the factors that control landsliding (Nowicki Jessee et al., 2018; Robinson et al., 2017) are hampered by some data availability issues. The EQTL grouping near ridge ridges and secondary clusters in the coludial slopes and above the inner quays was observed by Meunier et al. (2008) for the Northridge, Chi-Chi 1999 and 1993 earthquakes. The topography-controlled hot spots were also identified in the patterns of landslides triggered by the earthquakes in El Salvador in 2001, 2008 Wenchuan, 2010 Haiti, 2010-2011 Canterbury, 2013 Lushan and 2017 Jiuzhaigou (Evans & Bent, 2004; Fan, Scaringi, et al., 2018; Hough et al., 2010; Li et al., 2018; Massey et al., 2018; Xu, Xu, Shen, et al., 2014; Xu, Xu, Shyu, Gao, et al., 2015). The significantly uneven distribution of EQTL along the slopes differs from that of landslides triggered by precipitation, which tend to be distributed more evenly along different parts of the slopes (Meunier et al., 2008), as they are mainly controlled by precipitation patterns and the coupling between slope geometry hydromechanical properties of slope material. The response of a slope to the dynamic stress induced by seismic waves is the result of a complex interaction between the frequency and energy content of seismic waves, which depend on the source route-specific attenuation and local site conditions (material, layered, topography, etc.) that may amplify or amplify shaking at specific wavelengths (Massey et al., 2017; Meunier et al., 2008; Sepúlveda et al., 2005). Several studies investigated the relative contributions of the source, route and local effects on site-specific amplification phenomena (Ashford & Sitar, 2002; Del Gaudio & Wasowski, 2011; Gischig et al., 2015; Jibson, 2011; Moore et al., 2011; Rizzitano et al., 2014; Strink & Wartman, 2011). However, no factors were found to be responsible for the grouping of landslides; on the other hand, the various combinations of factors mentioned above work for different cases (Massey et al., 2017). Site-specific resonance (i.e. amplification) may occur through constructive interference of seismic waves. Reflected, refracted and/or diffracted waves, at the free surface or at the interfaces between geological structures with strong contrast of imedance, may contribute to resonance phenomena. The shape and orientation of these surfaces may cause amplification at multiple frequencies, depending on the topography and the thickness and lateral extent of the geological structures (Del Gaudio et al., 2014). Strong contrasts of impenence that produce local amplification may be caused by a layer of soil or debris above a rock formation, highly fractured areas within more intact materials or eroded materials that de-waste less eroded materials (Bouredeau & Havenith, 2008; Bozzano et al., 2008; Del Gaudio & Wasowski, 2011; Gischig et al., 2015; Moore et al., 2011). Site-dependent anisotropy may also occur, consisting of a pronounced resonance anisotropy that causes a significant amplification of agitation in a preferential direction (Del Gaudio & Wasowski, 2007). Surface morphology (tilt, convexity and height) can cause amplification both at the larger scales of the ridge and at the smaller scales of the site (Hough et al., 2010; Kaiser et al., 2014; Meunier et al., 2008). Several studies have shown amplification phenomena on sliding-prone slopes (Del Gaudio & Wasowski, 2007, 2011; Hartzell et al., 2017; Moore et al., 2011). These papers show that the amplification observed at different stations indicates peaks that coincide with potential sliding directions, which implies a higher susceptibility of the slope to seismic insufficiency in terms of what can be estimated by conventional (pseudo-static) analyses of slope stability (Del Gaudio et al., 2014). Although accurate information on site amplification effects can be obtained by recording strong-moving events for a long period of time (Hartzell et al., 2017), the need for a very dense accelerometer monitoring network (e.g. multiple stations on each in earthquake-prone areas makes it unfeasible to recover this information and alternative methods have been used. After noticing that ground movements from 3-D earthquake simulations show that topographic curvature topographical amplification, Maufroy et al. (2015) proposed a methodology for estimating amplification by exploiting such a correlation. Rai et al. (2016) also suggested an empirical approach calibrated on a set of records from small and medium earthquakes in California. Their model is based on a parameter called relative elevation, which quantifies topography using the altitude of a site in relation to its surroundings and improves the soil movement mitigation model previously proposed by Chiou et al. (2010). Ambient noise analysis as a replacement for strong-motion accelerograms (Figure 9) is useful in assessing the topographical amplification of the site, for example, through the technique of the horizontal-vertical noise spectral ratio, first proposed by Nogoshi and Igarashi (1971) and Nakamura (1989). Its working principle is that a strong contrast of imedance between an artificial soft layer and a more rigid substrate results in an amplification of the horizontal components of seismic background noise at the same frequencies at which the amplification of shear waves reaches its maximum level (Del Gaudio et al., 2014). Consequently, a pronounced peak in the H/V spectral ratios (horizontal/vertical) at site-specific frequencies may reveal site-specific (and directional) resonance conditions and information can also be obtained on the S-wave speeds of artificial material (Burjánek et al., 2010; Danneels et al., 2008; Del Gaudio et al., 2008; Gallipoli & Mucciarelli, 2007; Hancox et al., 2002; Havevenith et al., 2002; Jongmans et al., 2009; Louie, 2001; Méric et al., 2007; Mryen et al., 2017; Nunziata et al., 2009; Ohori et al., 2002; Torgoev et al., 2013). The site-dependent directorate was first observed by Bonamassa and Vidale (1991) and De Spudich et al. (1996) and was linked, in some cases, to the presence of pre-existing landslides that caused strong material contrasts to defined interfaces (Rial, 1996; Xu et al., 1996). Since directional amplification cannot be predicted solely from the topography of the surface, ambient noise analysis is an important tool for gaining insight into the underground structure (Del Gaudio et al., 2014), estimating amplifications at which it may experience a slope during a seismic event and assessing the probability of slope failure under different conditions. Summary of experimental approaches (discussed in this section) to investigate the mechanism of failure, dynamics and kinematics of earthquake-triggered landslides. Shaking tables are experimental devices that can be used to investigate the response of model slopes, artificial structures or building components to seismic shaking or other vibration snow (Figure 9). They are particularly useful in studying eQTL initiation mechanisms for backward analysis of which did not have adequate monitoring (R. Huang et al., 2013) and for measuring the mechanical parameters of weak layers that favour failure (e.g. Podolskiy et al., 2015). In fact, limited information on the resistance properties of soils and rocks under dynamic force prevent the applicability of physically based advanced numerical models that explicitly explain seismic agitation. To overcome this limitation, laboratory equipment capable of testing soil and rock samples or assemblies of different sizes under dynamic conditions has become increasingly used (Lin & Wang, 2006; Wang & Sassa, 2009; Wartman et al., 2005; Wasowski et al., 2011). Among others, Lin and Wang (2006), Liu et al. (2013) and G. Fan, Zhang, et al. (2016) performed tests of trembling on rock slopes with different rigidities, strengths and orientations of the bedding. They reported a positive correlation between the amplification of seismic acceleration and the height of the model slopes, which may result from reduced stiffness along the height (Ambraseys, 2009; Dakoulas & Gazetaas, 1986). These studies also demonstrated the role of weak layers (slope-direction diving) and fractures in improving seismic displacement and increasing the risk of failure. Through the tests of shaking meals on the sand slopes, Katz and Aharonov (2006) demonstrated the role of material heterogeneity in defining the model of distribution of landslide dimensions. They observed that small and superficial landslides appear as slumps in fairly homogeneous and unconsolidated soil deposits. The size of these landslides is essentially controlled by the depth of the deposit over the base rock, which is generally heterogeneous due to the presence of layers, bedding planes, joints and fractures. This heterogeneity is partly responsible for the degradation of the law of power in natural distributions of landslide sizes. R. Huang et al. (2013) used a shaking table to study the failure mechanism of a specific landslide triggered by the 2008 Wenchuan earthquake. They reproduced the deformation pattern before the general failure and noticed that the location of the failure plane in the model was consistent with that of the maximum horizontal acceleration. Based on this, the authors developed a simple empirical model to predict the depth of landslides in the rocky slopes in the characteristics of seismic agitation and rocky material, with the appearance of deeper landslides associated with the presence of softer/weaker rocks (i.e. rocks with less elastic mode and less cohesion). Wartman et al. (2005) investigated how permanent slope movements are generated to verify the reliability of slope travel predictions obtained by approaching Newmark's sliding block (1965). They demonstrated that such an approach provides good first-order deformation estimates, after Wang and Lin (2011) later confirmed. However, these estimates provide too much movement for slopes that have small displacements, but they are better suited to those cases where the sliding resistance decomposes towards residual resistance (reached after large displacements). Y. -M. Hsieh, Lee, et al. (2011) (2011) the need for table shaking tests to check the reliability of sliding thresholds, which may be much higher than the actual ones, if obtained by static tilt tests. Although slope conditions are well controlled and different scenarios can be tested accurately, upscaling results obtained by shaking-table tests on meter-size slopes to the size of actual slopes can be problematic, since material behavior scale nonlinearly. To overcome this problem, a geotechnical centrifuge can be used (Brennan & Madabhushi, 2009; Liang & Knappett, 2017; Taboada-Urtusastegui et al., 2002; Zhang, 2018; Zhang et al., 2017). Spinning hundreds of times per minute, the centrifuge can impose large accelerations on the test slope, producing resistance patterns that are more similar to those found in natural slopes. The scaling of the metric quantities shall then be carried out in relation to the ratio between the centrifugal acceleration imposed in the test slope and the gravitational acceleration in the natural slope. A recent study in a centrifuge was reported by Zhang et al. (2017), which investigated the case of a slope made from a layer of loess that over a layer of muddy stone. They found the amplification of horizontal and vertical seismic acceleration at the slope ridge up to three times that at the base and demonstrated the role of contrast in the properties of the two materials in contact in the production of traction cracks at the ridge and shear area at contact. Jibson (2011) summarized the existing methods for assessing the seismic stability of the slopes and estimating the amount of permanent inertia movements. The methods are classified into three categories, each having specific strengths, weaknesses and applicability in different situations: (1) pseudostatic analysis, (2) stress deformation analysis and (3) permanent displacement analysis (Figure 9). Pseudostatic analyses were developed at the beginning of the history of soil mechanics (Terzaghi, 1950). In such analyses, a permanent body force representing seismic agitation is introduced into conventional calculations of static limit equilibrium. In general, only the horizontal component of seismic agitation is shaped, since the cumulative effect of vertical forces is assumed to be negligible. However, simulations and experiments (Huang et al., 2001; Yuan et al., 2015; Zhang et al., 2015) shows that vertical shaking can significantly affect results. In fact, in the presence of vertical acceleration, weak materials may fail under a lesser amount of shear stress due to reduced normal stress or traction failure (Podolskiy et al., 2010b). On the other hand, the use of a permanent and one-way force to represent seismic agitation in pseudostatic analyses significantly underestimates the seismic stability of a slope. Consequently, scaling coefficients should be to represent only a fraction of the PGA, which, in fact, is only effective for a very short period of time (Blake et al., 2002; 2002; 2011; Jibson & Michael, 2009; Kramer, 1996; Marcuson, 1981; Seeds, 1979; Terzaghi, 1950; Wieczorek et al., 1985). The use of stress deformation analyses began to spread in the 1960s, after the development of the finite element method. Originally designed for general engineering applications, the method was soon used to model static and dynamic slope deformities. Other modelling strategies followed, including the finite difference method (Mitchell & Griffiths, 1980), discontinuous deformation analysis (Shi & Goodman, 1985), discrete element analysis (Cundall, 1971), rigid arc-block method (Kawai, 1978) and limit element method (Brebbia & Wrobel, 1980). Over time, increasingly complex and computationalally intensive models (e.g. nonlinear and inelastic) (Boulanger & Ziotopoulou, 2012; Elgamal et al., 1990; Gerolymos & Gazetas, 2005; Griffiths & Prevost, 1988; Prevost, 1981; Ziotopoulou & Boulanger, 2016). However, the use of complex models is only practical if high-quality, spatially available input data and high-quality input. Thus, advanced stress deformation analyses are generally used only in strategic construction projects or to shape individual slopes affecting essential infrastructures or rescue lines. The application of such analyses in larger and generic regions remains impossible, as the availability of required input data is normally insufficient (Jibson, 2011). The analysis of permanent displacement was introduced by Newmark (1965), who conceptualized the coseismic landslides as individual rigid blocks moving on an inclined plane when a critical acceleration (return) is exceeded. This critical acceleration depends on the tilt of the sliding surface from the horizontal and the safety factor, the ratio of resistance to motor forces or moments downhill. The entry for the method is a real or synthetic accelerometer. The speed and time-shifting of the block can be achieved simply by successively integrating the portions of the accelerometer in which the critical acceleration is exceeded. It is assumed that the modeled landslide behaves like a rigid plastic body, without internal deformations, and several other hypotheses of the model are generally made for simplicity (Jibson, 2011). The Newmark method has inspired more sophisticated methods that are capable of shaping the dynamic response of the landslide material more realistically and producing more accurate travel estimates. These models can be classified in decoupled and coupled analyses. Decoupled analyses represent the internal deformability of the mass of landslides under seismic shaking (Lin & Whitman, 1986; Seed & Martin, 1966). The most widely used procedure was proposed by Makdisi and Seed and consists of two stages. First, a dynamic analysis of the slope is carried out assuming that there is no fault surface, and an average acceleration of the sliding mass is obtained by average accelerations calculated at several points of the slope. Slope. the resulting acceleration is considered as input for a Newmark analysis to assess permanent displacement. This analysis does not take into account the effects of sliding motion on soil movement and does not take into account amplification in different parts of the slope. Coupled models go beyond some of the limitations of decoupled analysis by sharing dynamic sliding mass response and permanent displacement. They are the most sophisticated form of analysis of sliding blocks, but also the most computationalally intensive. Bray and Travasarou (2007) developed a simplified approach that uses a nonlinear, fully coupled, sliding-block model to produce a semi-empirical relationship for predicting movement. Jibson et al. (2013) published a software package for the implementation of rigid, decoupled and coupled analyses. Modelling is not the actual displacements of landslides. Rather, they are an index of slope performance under seismic excitation, which must be interpreted carefully in relation to the likely effect on a potential landslide (Jibson, 2007; Jibson et al., 1998, 2000; Pradel et al., 2005; Rathje & Bray, 2000). In general, the displacement of the threshold can be determined according to the characteristics of the landslide material. For example, the movements evaluated in the range of 5-10 cm may correspond to the ground fissure and previously undistorted soils approaching the residual resistance state. A static stability analysis can thus provide useful information on postseismic slope stability (Jibson & Keefer, 1993). Identifying and estimating spatial distribution patterns and the likelihood of EQTLs occurring remains difficult. With regard to the development of empirically or physically based models to predict eQTL size statistics and spatial distributions, the following issues should be addressed in future research: Calibration of both physically and statistically based models is based on the availability of EQTL inventories. Although more EQTL inventories are available than before, the representativeness and completeness of inventories are still a concern. For example, Tanyas et al. (2017) showed that 50% of all affordable EQTL stocks are from Asia and only 5% are from South America. In addition, available stocks differ in quality and completeness (Tanyas et al., 2017), which may directly affect the reliability of hazard assessments (e.g. Pelricani & Spilotro, 2015). For both physically and statistically based models, the accessibility of the required layers of input parameters, especially globally, and the estimates of their uncertainties are an ongoing challenge. More efforts are needed to better understand the control factors and their interactions in order to be able to predict the distribution of coseismic terrain in the future. For this, it is also very important to consider whether control factors can be spatially represented so that can be used in a predictive model. Ground movement characterization is an important component of rapid assessment studies, and ShakeMap is the most commonly used tool to obtain the necessary parameters. However, ShakeMap does not currently directly take into account topographical amplification (Sepúlveda et al., 2005), the duration of the shaking (Jibson, 2011; Jibson et al., 2004) or the direction of the rupture (e.g. Gallen et al., 2017), all of these may be factors in the process of initiating the landslide (Jibson, 2011; Jibson et al., 2004). After being discussed in Section 2.2, experimental studies are useful for understanding the mechanism of initiation and failure of EQTLs, but how to resolve the scale effect and apply the results for estimating EQTLs for a large area still requires more effort. In particular, the shear resistance of soils and rocks strongly influences slope stability, but none of the current available methods can estimate the strengths of shear and their relative contribution to susceptibility to regional landslides (Dreyfus et al., 2013). One of the immediate post-seismic hazards is the landslides that dam rivers, forming lakes that flood upstream areas and dams that could fail and catastrophically release the stored volumes of water and sediment (Figure 1). These landslide dams (Costa & Schuster, 1988) are common during or shortly after strong earthquakes (Evans et al., 2011; Keefer, 1984; Korup, 2004b; Schuster et al., 1998). Their lifespan varies from a few hours to thousands of years. Landslide dams are a well-known hazard given their potential to release burst floods; they usually require mitigation and evacuation of emergency engineering (Delaney & Evans, 2015; Fan, van Westen, et al., 2012; O'Connor & Costa, 2004; Peng & Zhang, 2012b, 2013b; Shi et al., 2017). Landslide dams can alleviate or accentuate water and sediment flows, channel geometry and stability and valley evolution (Hewitt, 2006; Korup et al., 2010). For example, Fan, van Westen, et al. (2012) estimated that landslide dams induced by the 2008 Wenchuan earthquake temporarily stored about 0.6 km3 of water and sediment. Both the gradual and catastrophic gaps in landslide dams release sediment downstream, storing materials on fans, months and delts (Hewitt et al., 2008). The degradation of the canal several to several tens of meters due to such increases in sediment yield was reported after major earthquakes (e.g. Yanetes et al., 2010). Lakes that have persisted for hundreds of years can alleviate flood waves, but even broken dams can control river morphology (Burchsted & Daniels, 2014; Hermanns et al., 2011; Hewitt et al., 2008), in particular by blocking local base levels (Korup, 2005; Korup & Tweed, 2007) or by creating armored frills of boulders that can remain in a river for thousands of years (eg, Korup, 2006a, 2006b). Groups of long-lived dams could suggest a seismic origin (Strom, 2015), and lake sediments may (Avsar et al., 2014, 2016; Howarth et al., 2012, 2014, 2016; Moernaut et al., 2014; Waldmann et al., 2011). Additional sedimentation discussions are provided for in Section 5. Despite several reviews (Costa & Schuster, 1988; Evans et al., 2011; Hewitt et al., 2008; Korup, 2002), some systematic analyses focused on earthquake-induced landslide dams; however, case studies abound (Dai et al., 2005; Dong et al., 2011; Fan, Xu, van Westen, et al., 2017; Schneider, 2009). We suspect that more scientific and practical information can be obtained from the study of earthquake-induced landslide dams to better inform risk assessment and emergency response. Schuster et al. (1998) listed four controls of the pattern of landslide dams generated by the earthquake: (1) seismic intensity (i.e. high PGA and duration of strong shaking); (2) steep hills and uneven topography; (3) weak and eroded litologists; and (4) high soil moisture or pore pressures. In addition to these controls applicable to landslides in general, the volume and mobility of landslides, the topography of the valley floor and hydrological factors (e.g. distance to the river network) determine whether landslides form dams (Fan et al., 2014; Tacconi Stefanelli et al., 2016). Predicting the place, time and manner in which a landslide forms a dam remains problematic, and several studies have focused on monitoring or quantifying the stability of landslide dams. Regional models and controls of coseismic landslide dams received less research attention compared to EQTLs. Most inventories of landslide dams refer to historical cases, although generally with unknown triggers and little data on location, size, type of landslides or damage caused. For example, the triggering of more than half of the 232 landslide dams examined in New Zealand remains unknown (Korup, 2004a). After the Wenchuan earthquake, Fan, van Westen, et al. (2012) compiled the largest inventory of event-based landslides (n = 828) to date and argued that the river dam trend depended on the type of landslide and initiation volume, as well as the average terrain relief and steepness (Fan et al. , 2014). Additional controls of candidates include distance from rivers, river power index and drainage density (Tang et al., 2018). Physically based models, such as the hydrodynamics of smooth particles and methods of discrete elements, can provide information on how landslide dams are formed (Braun et al., 2018; Zhao et al., 2017). A lot of effort is about understanding and predicting the stability and longevity of landslide dams. works explored several geomorphic predictors separating stability from unstable dams (Tacconi Stefanelli et al., 2016), although the distinction between intact and failed dams may be more appropriate. Some candidate predictors are the volume of the landslide dam (Vd), height, width, length, lake volume and upstream river basin (Ac; Casagli & Ermini, 1999; Dong et al., 2009, 2011; Ermini & Casagli, 2003). A block ingt, (3) (3) revealed several initial trends in landslide stability in the Northern Apennines (Casagli & Ermini, 1999); No land slip dams have been reported for BI < 3, while known landslide dams have remained intact for hundreds of years or more for BI > 5. Such empirical thresholds are likely to vary, depending on the area of study and the quality of the data. X. Fan, van Westen, et al. (2012) found that following the Wenchuan earthquake, no landslide dams were formed where BI < 2, while most dams were still intact about a year after the earthquake in which BI > 5. The internal structure of the landslide dam, the properties of the materials, the distribution of the grain size, the volume and speed of entry of water and sediments and the infiltration processes affect the stability and longevity of the landslide dams (Korup & Tweed, 2007). Unfortunately, the internal structure is exposed especially after dam failure, although geophysical measurements and geotechnical tests have begun to elucidate some internal characteristics of the selected dams (Casagli et al., 2003; Dunning & Armitage, 2011; Wang et al., 2013; G. Wang, Furuya, et al., 2016; Weidinger, 2011; Weidinger et al., 2002). Revisions of land landslide sedimentology, in particular landslides and rock avalanches, have contributed important details on the distribution of grain size, resistance of undercut deposits or possible water percolation paths (e.g. Dufresne et al., 2016). Landslide dams can fail through a scalding, ducting and slope failure (Costa & Schuster, 1988). Overtopping is the most common way (Schuster, 1993; L. Zhang, Peng, et al., 2016), with water spilling over the ridge of the dam and eroding a canal through the dam (Manville, 2001). Tube means internal erosion by infiltration, which removes solid particles and produces underground tubular pipes that initially appear as springs or infiltrations on the front of the downstream dam (Singh, 1996). Slope failure is initiated when the hydraulic pressure of the sequestered water exceeds the shear resistance of the dam material (Walder & O'Connor, 1997) and usually occurs with pipes and overgrounding when vertical erosion exceeds the lateral walls of the breach. The mechanisms of violation of the dam are less well understood and usually escape direct observation. Many dams have been artificially breached to avoid uncontrolled flooding (Canuti et al., 1999; Fan, van Westen, et al., 2012; N. Liu, Chen, et al., 2010; Xu et al., 2009); thus, the natural mechanisms of violation of the dam could not be observed. Field-scale dam breakage experiments have been conducted to understand the faultmechanism of landslide dams (Yan et al., 2017; Zhou et al., 2013). broadband can capture the characteristic seismic signals generated by flooding, debris flow and bed loading flow and thus help determine the timing and extent of dam failure and subsequent sediment transport (e.g. Chao et al., 2015; Feng, 2012; Yan et al., 2017). Increased land slip rates are common in earthquake-affected regions. Rainfall can trigger new landslides (Marc et al., al., of rock material weakened and damaged by shaking or remobilization of EQTL material (Domènech et al., 2019; Fan et al., 2019; Fan, Domènech, et al., 2018) which can feed debris streams for many years after the earthquake (Figure 1; Lin et al., 2004; Tang et al., 2009). However, our knowledge of the timing and extent of post-seismic landslides, debris flows and sediment transport depends on observations rather than predictions. This uncertainty makes it difficult to plan for reconstruction, especially for potentially damaged valley floors. Increased rates of geomorphic activity after large earthquakes were first described for the 1897 Assam earthquake. Oldham (1899) gathered some of the earliest accounts of landslides and post-seismic debris flows. With the exception of a pioneering study of the 1929 Murcison earthquake, New Zealand (Pears & Watson, 1986), post-seismic geohazards remained largely unstudied until the 1999 earthquake in Chi-Chi, Taiwan (e.g. Hovius et al., 2000). Compared to previous rates of land slip flows and sediment flows, the 1999 Chi-Chi earthquake resulted in a rapid increase in suspended sediment flows in rivers (Dadson et al., 2004), followed by an increase in land slip rates in 2001 Typhoon Toraji (Hovius et al., 2011). Subsequent analyses suggested that improved land slip and sediment flow rates, when normalized for climate records, can be maintained for a decade (Marc et al., 2015). Increased landsliding rates immediately after earthquakes may not only be driven by rainfall or aftershocks, but also by reduced strength of hillslope materials due to intense mainshock shaking. Estimates of post-seismic sliding rates come from detailed compilations of inventories collected from satellite images (e.g. Fan et al., 2019; Marc et al., 2015), and these studies show that the pulses of increased frequency of landslides and redness of sediments occurred within 1-10 years of the earthquake. Many studies have reported that both landsliding rates and degradation transport sediments toward fund rates within a decade (Dadson et al., 2004; Hovius et al., 2011; Khattak et al., 2010; C. -W. Lin, Liu, et al., 2006; Marc et al., 2015; Shou et al., 2011; Yang et al., 2017; S. Zhang, Zhang, et al., 2016; Zhang & Zhang, 2017; Figure 10). Normalized landslide rates in terms of climate (solid circles, left axis) and precipitation that triggers debris flow (open circles, right axis) before and after some recent earthquakes. Landslide rates are the ratio of average landslide volumes before the earthquake to landslide volumes for given time. The scaling of the area volume proposed by Laren et al. (2010) for mixed soil and landslide data sets is used here to estimate the volumes in the total landslide areas (although other options are possible). Vertical bars indicate the error (standard deviation ±1) associated with scaling. Climate normalisation was achieved by dividing volumes by cumulative rainfall between the date of previous teledetection image and given time. The size of the open circles (precipitation that triggers the flow of debris) is proportional to the number of debris streams triggered by the respective precipitationats at that time. Data from Domènech et al. (2018); Fan, Domènech, et al. (2018); Fan et al. (2019); Khattak et al. (2010); Marc et al. (2015); and Saba et al. (2010). Detailed mapping of 34 time slices between 1994 and 2006 and therefore before and after the 1999 Chi-Chi earthquake argued the importance of rock deterioration in relation to climate controls and seismic force (Dadson et al., 2004; Marc et al., 2015; Figure 10). Land slip rates V correlated with total precipitation P exponentially: (4)where A and B are empirical constants reflecting pre-earthquake susceptibility to slope failure in the studied area and geomorphic efficacy of precipitation (Marc et al., 2015). In regions affected by the 2008 Wenchuan earthquake, post-seismic landliding growth rates were higher than previously observed, though less than a decade (Fan, Domènech, et al., 2018; Tang et al., 2016; Yang et al., 2017; S. Zhang, Zhang, et al., 2016). However, all inventories document intense and short periods of landslides and post-seismic debris streams, which has implications for the overall budget of earthquake sediments (see section 5). For example, the pulse of post-seismic sediments associated with the 2008 Wenchuan earthquake mobilized 0.6–1.5 km3 of material. The volume of active landslides decreased by >70% over 3 years and virtually normalized within 7 years of the shock, even though most sediments remained in place. This suggests that the legacy of sediments generated by earthquakes could last, in various forms, much longer than the decade after the earthquake (Fan, Domènech, et al., 2018). The extent and frequency of reported post-seismic malfunctions differ from one study area to one. Post-seismic-sized distributions of landslides for the 1993, Chi-Chi and 2008 iwate earthquakes were indistinguishable from the inventories of coseismic landslides (Marc et al., 2015), and the authors deduced a uniform decrease in ground resistance at shallow depths during the earthquake. This is followed by a recovery of ground resistance, independent of the size or depth of the malfunction. Contrasting observations were reported during post-seismic remobilizations of coseismic landslides triggered by the 2008 Wenchuan earthquake (Fan, Domènech, et al., 2018). Although smaller coseismic landslides quickly stabilized, the larger and deepest ones remained prone to reactivation a few years after the earthquake. The form of the distribution of the frequency of landslides and post-seismic debris flows, which reaches its peak following a and degrades to pre-earthquake values over years, is consistent among different earthquakes (Figure 10). Some authors argue that we can see the mechanisms that control the pattern of landslides before and after an earthquake from changes in their size In terms of remobilisation of landslides, debris flows appear to play a major role. Active landslides tend to throw debris streams along narrow gullies, partly through the hortonian land flow that causes sedimenttotrain (e.g. Domènech et al., 2019; van Asch et al., 2014) or collapse of saturated water material (Iverson, 1997). Experiments on artificial slopes using debris from coseismic landslides provide an insight into how changes in the size distribution of landslides affect the initiation and depletion of flow-like landslides (Hu et al., 2017; Hu, Hicher, et al., 2018; Hu, Scaringi, et al., 2018; Wang & Sassa, 2001, 2003). These studies have highlighted that the coarseness of hydrologically operated cereals, for example through artificial and internal erosion and depletion of fine particles, can improve the stability of deposits by increasing hydraulic conductivity and intergranular contacts, reducing the likelihood of failure in larger rain storms (Hu et al., 2017; Hu, Hicher, et al., 2018). Studies monitoring post-seismic debris flows have reported a trend towards coarser flows that have travelled shorter distances and been initiated in larger EQTL deposits (e.g. Chen et al., 2014; Zhang et al., 2013). These coarser scrap streams indicate either reworking or winnowing of fines from landslide debris or more storage of coarse material in channel beds. The progressive depletion of erodable debris, increased hydraulic conductivity of the soil, increased sediment friction due to compaction and increased root consolidation in the coverage of the reguroated vegetation could explain the degradation of the activity of debris flow in the years after the earthquake (e.g. Domènech et al., 2019; Hu, Scaringi, et al., 2018; Z. Li, Jiao, et al., 2014; Saito et al., 2014; Shen et al., 2017; Yang et al., 2018). Exhaustion of the total volume of available sediments (Cruden & Hu, 1993) is unlikely because most EQTL sediments (up to 80-90%) can remain on the hillslope for many years after the initial event. Landslides that remain active after the earthquake tend to occur closer to the canals and in some parts of the landscape with larger drainage areas (Fan et al., 2019) than latent landslide deposits, although remobilization could not depend solely on river undercutting (e.g. Croissant, Lague, Davy, et al., 2017; Croissant, Lague, Steer, & Davy, 2017) where flow power is low and coarse sedimentprotects the base of landslide deposits. Several hypotheses relate to the reduction of post-seismic landslide activity. Firstly, changes in connectivity landslide within the river or hill hydrological system could alter the thresholds for triggering the flow of debris (Z. Li, Jiao, et al., 2014, see section 5). The debris incision spills into large land sliding masses can disrupt the flow of groundwater, which can lead to shallow basement drainage and reduced the potential for high pore pressures needed to trigger debris flows. Secondly, extensive mitigation work can mitigate the scrap flows (e.g. Liu et al., 2017) and control dams and other restraint systems limit the export of sediment from low-order river basins. Thirdly, vegetation can rapidly recover the slopes of earthquakes, especially in wet tropical and temperate regions, thereby reducing the land flow and stabilizing the hill slopes by anchoring the roots (e.g. O'Loughlin, 1985;

Schmidt et al., 2003; Watson et al., 1999) and finally closing potential sediment sources for debris flows (Y. Liu, Liu, & Ge, 2010). Fourthly, the coarseness of cereals resulting from the progressive gain of fines or recurrent land (Hu, Hicher, et al., 2018; Hu, Scaringi, et al., 2018; S. Zhang, Zhang, & Chen, 2014) increases hydraulic conductivity, making surface leakage and excessive pore pressure less likely. A measured increase in suspended sediment concentrations (grain size < 0.25–0.5 mm) in rivers shortly after the Wenchuan earthquake, followed by degradation in subsequent years (Li et al., 2017; J. Wang, Jin, et al., 2015; Wang et al., 2017), could be evidence of this rough ingtreor process or simply reflect artifacts of debris flow (S. Zhang et al., 2013). Post-seismic landslides may form at higher rates during heavy rains, possibly caused by the deforestation of over-secured slopes in EQTL crowns, vegetation and groundwater changes, and modified mechanical properties of base rock and regolith (Marc et al., 2015). The overread slopes of the crowns of coseismic landslides contributed less than half of the total volume of postseismic landslides in documented cases (Schöpa et al., 2018). Vegetation and groundwater changes have different time intervals than transient post-seismic landslides (<10 years). The reduction of earthquake-induced power can occur through various mechanisms in colluvium, regolith and base rock. In colluvium and regoligoth, mechanical damage by decompression and expansion and dilation of superficial cracks could affect the rate of triggering of landslides. Base rock cracks, joints, and bedding planes can form or widen during shaking. The emergence of landslides with a number of depths suggests that these different mechanisms produce a similar geomorphic response. These changes in the properties of underground materials appear to be transient and can only be exploited during extreme precipitation events, as landslide rates remain low in intervals without extreme precipitation (Marc et al., 2015). It is difficult to validate these proposed changes in the shear resistance of the slope material of the hills, since direct geotechnical testing is based on volumes and obtaining a statistically significant amount of data is too costly and time-consuming. One possible solution is to use the characteristic speeds of seismic waves, which depend on the elastic module of the environment in which they propagate. Laboratory experiments found a reduction in elastic modulations caused by moderate strains in rock samples and subsequent logarithmic recovery or healing (Johnson et al., 1996; Rivière Rivière al., 2015). Similar results were obtained during ground shaking field experiments (Lawrence et al., 2009; Renaud et al., 2014). Interferometry and ambient noise techniques analysing seismic waves by deconvolution revealed changes in underground seismic speeds at depths of 100 m to several kilometres after medium to large earthquakes (e.g. Brenguier et al., 2008; Hobiger et al., 2014). In these cases, the decrease in the speed due to the earthquake is constantly followed by a progressive, non-linear recovery over the months to years. Large-scale damage to the substrate was attributed to strong soil movement and failure, and post-seismic recovery was attributed to the healing of grain cracks and contacts (Sawazaki & Snieder, 2013). These mechanisms could also affect the strength of the hillslope material. Therefore, studying the coevolution of epicentral landslide rates and shallow seismic velocities could go in a certain direction towards explaining the transients observed in the waste of the post-seismic mass, at least that the detachment of the base rock, since the soils and mantles of debris remain unaffected by changes in the power of the mass of the bulk rocks. Rainfall is a common trigger for landslides (e.g. Crinse & Frattini, 2001; Froude & Petley, 2018; Segoni et al., 2018), as well as processes related to precipitation at the appearance and size of slope failure, any remobilization, acceleration and fluidization have been extensively studied (e.g. Crozier & Blade, 1999; Jamei et al., 2017; Montgomery & Dietrich, 1994; Olivares & Damiano, 2007; Wieczorek et al., 2000). However, we can still learn more about the variability and scaling of meteorological and hydrological factors and the processes that control slope failure (Almeida et al., 2017; Won & Germann, 2013; Corominas et al., 2013; Chita & Frattini, 2001; von Ruette et al., 2014). Technical advances now allow very detailed measurements of hydromechanical changes in a landslide body (Kogure & Okuda, 2018). Shallow landslides and debris flows are more directly related to patterns of precipitation intensity, duration and previous rainfall (e.g. Aletti, 2004; Guzzetti et al., 2008). Precipitation thresholds are a condition which, when exceeded, is likely to trigger landslides (Guzzetti et al., 2008). The thresholds may be lower limits of known hydrological conditions (e.g. precipitation, infiltration and soil moisture) which triggered landslides or upper limits of the same conditions which did not trigger landslides (Reichenbach et al., 1998; Staley et al., 2013). The spread between these thresholds is usually large or poorly delineated. Simple empirical relationships could be as it neglects important control factors (such as variability of topography and soil types) and should be best reported likely if used in early warning and risk reduction systems (e.g. Basher et al., 2015; Piiullo et al., 2017; Rosi et al., 2012; Vennari et al., 2014). Practical thresholds have already been incorporated into regional early warning systems (e.g. Bogaard & Greco, 2018; Segoni et al., 2018, 2017), see also section S4.3 for an overview of applications from different countries. Precipitation thresholds may represent changes in the hydrological and geotechnical conditions of the slopes, including the effects of earthquake-induced slope weakening, weather, vegetation fires, climate, land use changes, slope consolidation processes and soil cover depletion (e.g. Brooks et al., 2002; Cannon et al., 2008; Fan, Juang, et al., 2018; Gariano et al., 2017; Kean et al., 2011; C. -W. Lin, Liu, et al., 2006; Shieh et al., 2009). The ubiquitous landslide sliding debris after the chi-chi and wenchuane earthquakes of 1999 and wenchuan in 2008 (see sections 3.2.1–3.2.3) was considered an important factor for the decrease in rainfall thresholds and the increase in the frequency of remobilization of landslides and debris flows (e.g. H. Chen & Hawkins, 2009; Huang & Li, 2014; Lin et al., 2004; Shieh et al., 2009; Tang et al., 2016, 2009; Yu et al., 2014; S. Zhang & Zhang, 2017). In an area affected by the 1999 Chi-Chi, Lin et al. (2004) and C.- W. Chang, Lin and Tsai (2011) found that the intensity of precipitation triggering debris flow and total precipitation decreased significantly, while large debris flows became more common in smaller river basins. After the 2008 Wenchuan earthquake, more frequent and destructive debris flows also occurred in the earthquake-affected region, and precipitation thresholds for triggering debris flow were significantly reduced (e.g. Yu et al., 2014; Zhou & Tang, 2014), up to 75% (Zhang & Zhang, 2017; Figure 11). They also reported similar reductions of 60% after the Chi-Chi earthquake and 50% after the 2005 Kashmir earthquake (M 7.6). These critical precipitation thresholds are gradually returned to pre-earthquake values (Figures 10 and 11). How long it takes for this recovery is still an open question, and estimates range from less than a decade to as long as a century (Cui et al., 2009; Huang & Fan, 2013; Koi et al., 2008; Yu et al., 2014; Zhang & Zhang, 2017). Changes in precipitation thresholds for debris flow after the Chi-Chi earthquakes of 1999 and Wenchuan in 2008. The maximum intensity of precipitation per hour during the precipitation event that triggered the debris flows is plotted. The data for the Chi-Chi earthquake are from Lin et al. (2004) and refer to the River Chenyulan river basin; the data for the Wenchuan earthquake are from S. Zhang and Zhang (2017) and refer to the epicentral area from Yingxiu to Gengda. Earthquakes can initiate, propagate, widen or interconnect cracks in rock exposures at a depth, especially near ridge ridges and high on slopes where shaking is improved by topographical effects (see section 2.2.1). Cracks facilitate weathering through cycles of heat stress and freezing, oxidation and hydrolysis, humidity and infiltration of electrolyte solutions (e.g. Ayonghe et al., 1999; Price, 1995; Scott et al., 2016; Whaling, 1984). Thus, the failures of the slope conditioned by such weather could happen for several decades earthquake (e.g. Towhatha, 2013). Post-seismic landslides caused by long-term rock degradation have been reported from Japan (Okamoto et al., 2013), and precursor earthquakes could play a role in strengthening the landslide triggered by the next earthquake (Parker et al., 2015; Wei et al., 2014). A recent example of a destructive landslide on a possible slope pre-conditioned by an earthquake is that of the 2017 Xinmo landslide, Sichuan, China, which mobilized several million cubic meters of rock from a mountain ridge and destroyed an entire village, killing many inhabitants (Fan, Xu, Scaringi, et al., 2017; Fan, Xu, & Scaringi, 2018; Scaringi et al., 2018). The slope was intensely rocked by the 1933 earthquake at Diexi (M 7.5), which also triggered large landslides nearby. Some of them produced landslide lakes, such as Lake Diexi (Cheng et al., 2008; Ling, 2015). Although the predisposing action of the 1933 earthquake on the Xinmo landslide has not been proven beyond any doubt, there is the possibility. Signs of incipient landslide were unnoticed, but further analysis of optical and satellite radar images revealed several interconnected cracks near the mountain ridge for more than a decade, and slow deformations occurred in the months prior to the event (Fan, Xu, Scaringi, et al., 2017). Five to ten thousand people die from landslides every year in the world (Froude & Petley, 2018; Petley, 2012), with a trend that is increasing over time, possibly due to human activity and global population growth. An unknown number of these occurred in seismic regions and could have delayed post-seismic landslides. Fan, Xu and Scaringi (2018) argued that high-altitude mountain ridges in inhabited seismic regions should be systematically scanned for structural features (e.g. extensive cracks and slow deformations) that can predict rock damage or imminent failure, so that real-time monitoring, warning systems and other countermeasures can be used to reduce the risk of such failures. However, delayed post-seismic landslides remain a poorly explored research direction, primarily because they are difficult to recognize. Few studies have attempted to quantify the potential for damage to the basement based on the size of earthquakes, nor have studies quantified how earthquake damage contributes to the production and transport of regolith from the hills. Seismic reflection or acoustic emissiondata data could help estimate rock damage (Clarke & Burbank, 2011; Von Voigtlander et al., 2018) and thus assess the post-seismic responses of rock mass properties. Despite the increasing number of inventories of coseismic landslides (see section 2.1), detailed data and time series of geomorphic reworking of the materials thus produced are few (X. Fan et al., 2019). Future research may want to consider the following: The causes for peak landslide rates after earthquakes remain insufficiently understood. Although Although remobilization of coseismic deposits and their transformation into flow-like landslides are now tangible at low complexity and simplified laboratory experiments, more efforts are needed to expose this understanding to the hills or river basins in a way that allows predictions and overcomes the limitations of threshold-based empirical approaches. The decade-long effects of earthquakes on slope stability and landslide rates require more research. More systematic studies of the time-based effects of earthquakes on soil and rock resistance and slope degradation and monitoring in seismic zones could help to identify less obvious defects in the postseismic slopes. Observations from regions hit by earthquakes in recent decades or centuries can also be beneficial to understand what to expect on mountain slopes affected by recent earthquakes. The forecasting of the potential of landslide dams has been helped by increased stocks. These observations led to the evolution of empirical patterns of landslide formation. However, physically based insights about the tendency to, and longevity, landslide dams are still in its infancy. Although many studies have examined the sedimentology of rock avalanches and rock slides (Dufresne et al., 2016; Dufresne & Dunning, 2017; Dunning & Armitage, 2011; Weidinger et al., 2014), hydrological and meteorological measurements are limited, in situ monitoring was rare, and the geotechnical characteristics of dam materials (e.g. particle size distribution, porosity, hydraulic conductivity and mechanical properties) require better study. Landslide susceptibility maps list the terrain in areas that have a different vulnerability to slope failure due to different combinations of causal factors. Landslide hazard maps also quantify the probabilities (spatial, temporal and intensity probabilities) for these areas (Corominas et al., 2013). Susceptibility maps are the most ubiquitous and are generated in cases where probabilistic assessments are not feasible, for example, where landslide data do not have time stamps. Landslide hazard maps may display different types of information, depending on the method with which they were generated and the historical data available. The frequency of landslides is generally derived from the frequency of the trigger (e.g. earthquake or precipitation). Landslide hazard maps that have been generated by statistical or heuristic methods may display the expected density of landslides as spatial-temporal probability (densities for specific frequencies). Maps mass generated by initiating and shaping the physical-based runout may show the intensity of the landslide (such as safety factor, probability of failure, impact pressure, deposit depth or flow speed) for specified scenarios or return periods (e.g. Jibson & Michael, 2009). Risk is defined as the likelihood of harmful consequences and therefore expected losses, e.g. death, injury, property, livelihoods, disrupted economic activity or environmental damage. This This interaction between hazards and vulnerable conditions (UNISDR, 2017). In order to estimate the risk, several important components are needed: the hazard (probability and temporal and spatial intensity), the quantification and characteristics of the elements exposed-to risk (e.g. buildings, persons and roads) and the physical vulnerability of these elements (which relate to the intensity of the hazard to the degree of deterioration for certain types of elements-at risk). If the different components of the risk equation can be spatially quantified for a specific set of hazard scenarios (HS) and risk elements, the risk can be analysed using the following equation (van Westen et al., 2017): (5) where P(T.HS) is the temporal probability of a particular SH. A SH is a hazard event of a certain type (e.g. , a landslide of a specified magnitude and recurrence interval); P(S? HS) is the spatial probability that a particular location will be affected, given a specific SH (e.g. landslide density for a particular susceptibility class). In the case of hazard intensity maps obtained by physical modelling, the spatial probability shall be considered to be 1; A(ER-HS) is the quantification of the number of items exposed to risk, taking into account a specific SH (e.g. number of persons, number of buildings, monetary values and hectares of land); and V(ER? HS) is the vulnerability (as a value between 0 and 1) of the elements at risk, given the intensity of the hazard within the specific SH. Estimating the sensitivity and post-seismic and post-seismic hazard, much less the risk, is therefore difficult, as shown in Figure 12, showing the expected changes in active landslides, exposed elements and methods used to generate landslide inventories, sensitivity, hazards and risk after a major earthquake that triggered landslides. Summary of expected changes in active landslides and risk-exposed elements, as well as the methods used to generate landslide inventories and assess sensitivity, hazards and risks after a major earthquake triggering landslides. (a) Schematic representation of the number of landslides before and after the earthquake. The above arrows indicate trigger events (red earthquakes and blue precipitation events), where the size of the arrows indicates different magnitudes. The earthquake triggers landslides, some of which could hit the ground with rivers. After the earthquake, landslide dams rapidly decompose as they are breached or stabilized, and the number of active landslides triggered by the earthquake gradually decreases, but extreme precipitation events could reactivate some of them (light red) it could also trigger new landslides (blue); (b) due to damage to the earthquake, the number of items presenting a risk (e.g. buildings) decreases drastically and begins to increase after the event, depending on the speed of recovery. Extreme precipitation phenomena can lead to debris streams and landslides, which can damage the reconstructed infrastructure, leading to a saw tooth pattern; (c) methods of inventorying landslides consisting of object-based image analysis and visual interpretation, depending on the availability of high-resolution images without clouds (red arrows represent for earthquake-triggered landslides and blue arrows represent for post-seismic debris flows); (d) methods for assessing the sensitivity to landslides. This should be constantly updated after the occurrence of the earthquake and subsequent precipitation events. Immediately after the earthquake, a rapid assessment of landslide density is required for rapid response planning before the landslide inventories are generation; (e) hazard assessment methods before, immediately after and in the years following the earthquake; (f) the assessment of the risk of landslides before, immediately after and in the years following the earthquake, leading to a steady decrease in risk as a result of the recovery of the natural and built environment. QRA is the quantitative risk assessment; PAGER is the prompt assessment of global earthquakes for response (Earle et al., 2009). Both PAGER and Shakemap are produced by the U.S. Geological Survey Earthquake Hazard Program. We will illustrate the scope of a cascade of seismic risk by the chain of events that hit the city of Beichuan in six major episodes during and after the Wenchuan earthquake (Figure 13; L. M. Zhang, Zhang, & Huang, 2014). In episode I, strong shaking of the seismic soil caused the collapse of about 80% of the buildings in the city of Beichuan. In episode II, the large landslide in Wangjiayan and the Jingjiashan rock avalanche triggered by the shaking buried parts of the city, killing more than 2,300 people. A large landslide at Tangjiashan, 4.5 km upstream from the city, blocked the Jianjiang River, forming a landslide dam with a lake volume of 3.16 × 108 m3; two smaller landslide dams also formed just upstream of the city. In episode III, on 10 June 2008, the Tangjiashan landslide dam was overrun and the resulting flood wave overcame the two smaller dams (Fan, Tang, et al., 2012), increasing discharge and causing widespread damage in Beichuan; fortunately, residents were evacuated earlier (Peng & Zhang, 2013a, 2013b). In episode IV, from 23 to 25 September 2008, the debris of landslides around the city fed the debris streams that buried much of the Old Town, which remained unaffected by episodes II and III. The Jianjiang River, which runs through the city, also changed course after the flows of debris in September 2008. In episode V, the riverbed near the city increased by 15-38 m due to sedimentation in the 5 years from 2008 to 2013. Finally, in episode VI, about 80% of the city was flooded in July 2013; due to the significant increase in the a much smaller flood than that of 10 June 2008 caused more severe flooding. The deserted city of Beichuan now serves as a memorial to the tragic earthquake and the cascade of adverse effects it has unleashed. Six episodes of a chain of danger that affected the city of Beichuan, China, China, Wenchuan earthquake. Many landslides and debris streams following the Wenchuan earthquake remained lethal in the area for many years after the earthquake, indicating that hazard assessments must be extended beyond traditional practice. Only the filling of landslide lakes allowed sufficient warning times to allow the evacuation and installation of technical measures to minimise the impact. Risk management involves at least three questions: (1) What can go wrong? (2) What is the probability that this will go wrong? (3) What are the consequences if it goes wrong? The risk assessment of multiple hazards is generally based on independent analyses that adopt disparate procedures and resolutions, so that more problems arise with this strategy: separate investigations of individual processes can misassess the overall risk profile (Bell & Blade, 2012); Comparing risks of different origins is difficult (Marzocchi et al., 2012); The weighting of different hazards and vulnerability factors are instantaneous that require a periodic update; Taking independent sources of risk neglects possible interactions between hazards (Kappes et al., 2012 Marzocchi et al., 2012); Assuming that vulnerability is risk-independent neglects changes in resilience against subsequent hazards; aggregation of individual, independent risks could misunderstand the total risk. Quantifying the risk in the hazard chains thus remains difficult. Landslides evolve as a chain of disasters with unknown results, and hazards can be correlated. Uncertainties and feedback are part of the entire chain of dangers and justify more research (Kappes et al., 2012; Lacasse et al., 2012). Based on Beichuan's experience (Figure 13) and other studies, L. M. Zhang, Zhang and Huang (2014) and Zhang and Zhang (2017) propose a preliminary protocol for assessing the risks of a cascade of danger, highlighting risk feedback and the concept of a life cycle: The Life Cycle of Hazards. Each process in a danger cascade has a certain period of pronounced activity (Figure 13). Therefore, each hazard has a life cycle; for example, from the formation of a landslide dam to its failure or the time it takes for a pulse of sediment to pass a certain stretch of the river. The first step is to identify these life cycles. Duration. Starting from the earthquake, each individual hazard should be assessed throughout its entire life cycle. Able. Each individual hazard should be investigated on its potentially affected surface during its life cycle. For landslide dams, consider the maximum possible upstream flood zone and potential downstream flooding. Items in danger. Exposed Items should be identified in areas potentially covered by hazards during their life cycles (Table 4; Chen et al., 2016; L. M. Zhang, Zhang, & Huang, 2014). Table 4. Recommended procedure for risk assessment in key stages of hazard chain tasks What can go wrong? 1. Select likely extreme hazards beyond design levels as reference 2. Consider fully the interactions with cascading hazards and hazards. 3. Assessment of the evolution of an initial hazard over its life cycle. Danger events may not occur at the same time. What is the probability that it will go wrong? 1. Express possible interactions of hazards in space and time using event trees. 2. Assessment of the probabilities of occurrence of hazard events using physically or statistically based methods (Keefer, 1984; Liu et al., 2015; L. L. Zhang, Zhang, et al., 2011). What are the consequences if it goes wrong? 1. Evaluate the space in which each individual hazard evolves over its life cycle. 2. Consider possible cascading effects. 3. Identification of the elements at risk in the hazard-affected space during their life cycle. 4. Investigate possible interactions between vulnerabilities to individual hazards and the effect of early warnings. Table 5 summarizes the feasibility of quantifying the cascading risk of seismic risk before, immediately after, a few years after and decades after an earthquake. Table 5. Multihazard Relationships and Feasibility of Quantification of Risk Components Before an Earthquake Multihazard Phenomenon Spatial Probability Intensity of Physical Probability Elements of Physical Vulnerability-to-Risk Landslides induced by Landslides and Precipitation Records Physically based on Statistical/Heuristic Modeling If runout modeling is included, otherwise only density Often not available an appropriate intensity. Possibly for runout. They can be obtained from the classification of satellite or drone images, voluntary mapping through OpenStreetMap, detailed survey or existing databases. Specific attributes (type, use, value and persons) should be collected for risk assessment. OpenStreetMap or earthquake-induced landslide scan scan scans Many earthquakes scenario possible Density for scenarios only No available intensity measure Only consider the destruction of exposed flood-induced precipitation strips If available discharge/precipitation records Flood modeling if sufficiently detailed data are available on elevation terrain coverage and cross-sections of rivers. Connecting with the earthquake-induced-flood stage deterioration curves Difficult to quantify many chains Prediction of landslide locations is not possible Limited flood models by lack of parameters Connecting with stage deterioration curves Shortly after activity Coseismic landslide Not relevant: Event has already occurred rapidly modeling the density of landslides and analyzing and interpreting object-oriented images Using the inventory of landslides to characterize locations Rapid damage mapping voluntary geographic information (VG, e.g. Humanitarian OpenStreepMap), classification of satellite images, and drones. The failure of coseismic landslide dams Rapid assessment of the danger of dam failure Modelling of dam flooding scenarios is time-consuming. Mechanism to break the dam of uncertainty, the modeling of downstream floods requires high resolution DEM. Known but unknown unknown curves Rapid mapping of the new situation using VGI. Over several years after the break-up of the flood landslide dam Monitoring of lake level and influx Modeling flood scenarios, based on dam monitoring, downstream flood modeling requires a high resolution DEM Connecting the type of building with known curves Rapid change, several steps of mapping required, using the classification of satellite or drone images, voluntary mapping through OpenStreetMap and detailed study. Specific attributes (type, use, value and persons) should be collected for risk assessment. OpenStreetMap or Study Reactivation of Precipitation-Induced Landslides Changing Precipitation Thresholds Satellite or Drone Monitoring Estimated Volume and Runout Is Difficult Often no intensity measures are possible. Consider only the destruction of exposed elements Precipitation-induced debris flows Change rainfall thresholds Many river basins with the probability of debris flow Unknown volume in modeling Develop new curves or use V = 1 approach. Floods induced by precipitation due to mass movement activity Depend on many chains Location and blocking volume is very difficult to predict. Flood modeling for potential scenarios is possible, but it depends on damming scenarios. Connecting with stage deterioration curves Decades after precipitation-induced flooding in sediment-filled valleys Return periods based on previous records cannot be used Flood modeling using new EMDs with new sediments Linking with phase deterioration curves Updated data using satellite or drone image rydens, voluntary OpenStreet Map mapping and detailed study. Specific attributes (type, use, value and persons) should be collected for risk assessment. OpenStreetMap or study Landslides induced by precipitation in recovered land Depends on landslides and precipitation records Physically based, statistical/heuristic mödeling If runout modeling is included, otherwise only density Often no measure of intensity Landslides that can be triggered by the next earthquake Many earthquake scenario Learn from the last earthquake model No measure of intensity Only consider the destruction exposed elements Note. Red = low; Orange = moderate; Green = High; VGI = Geographical information provided voluntarily; DEM = digital elevation model. V = 1 indicates that the exposed elements are considered to be total destroyed (vulnerability = 1). See text for explanation. Estimating the sensitivity, danger and risk of EQTL requires knowledge of the possible locations of future earthquakes and the including magnitude, depth, length of rupture, presence of surface rupture and source mechanism (Marc, Hovius, & Meunier, 2016). This information is rarely known in advance, although a set of plausible earthquake scenarios can be identified that could produce different patterns of landslides in terms of fault types, size distributions and density. Nadim et al. (2006) used a 475-year PGA derived from the Global Seismic Risk Program as one of the triggers in a multifactor for estimating EQTL sensitivity worldwide. Such maps highlight possible regions of interest, but cannot be used for landscaping (Cepeda et al., 2013). Detailed quantitative methods for earthquake-induced risk assessment have been developed, based on the analysis of event trees, for specific installations with high potential, such as nuclear power plants or hydropower dams (Kwag & Hahm, 2018), but cannot be applied on large areas. The basis of many studies on EQTL sensitivity and hazard assessment is the Newmark model (1965), which involves slipping a rigid block on a plane inclined under the influence of seismic agitation (paragraph 2.2.3). Cumulative displacement of the downward slope induced by an increase in agitation is an index of the seismic stability of the slope (Jibson, 2007). Wilson and Keefer (1983) used strong motion footage and field observations of a landslide triggered by the 1979 Coyote Creek (California) earthquake to show that Newmark's model accurately predicts the dynamic behavior of landslides on natural slopes. Wilson and Keefer (1985) proposed a framework for using the Newmark model to produce seismic slope stability maps on a regional scale, and Wieczorek et al. (1985) applied this concept to an experimental seismic slope stability map. Jibson (1993) developed a simplified Newmark method for regional analysis by estimating Newmark displacement according to earthquake intensity and critical acceleration, i.e. the minimum ground acceleration required to initiate slope failure; Jibson (2007) updated this methodology using a much larger data set. Jibson et al. (1998, 2000) compared Newark-shaped displacements with landslide data and developed a simple mathematical relationship demonstrating that Newmark's estimated displacement correlates with the frequency of landslides, providing a simple, physically based approach to estimating seismic ground sliding hazards on a regional scale. Newmark-displacement models have been refined with data on the steepness of the local slope; geotechnical properties, such as cohesion, friction angle and density related to geological units; and groundwater levels (Del Gaudio et al., 2003; Hsieh & Lee, 2011; Jibson et al., 1998, 2000; Liu et al., 2018; Miles & Keefer, 2001; Wang & Lin, 2011; Y. Wang, Song, et al., 2016). Newmark displacement analysis can be based on deterministic or probabilistic approaches. Strong movement data are the main contribution for deterministic analyses (Biondi et al., 2004; Gallen et al., 2017; Rodriguez-Peces et al., 2011; Romeo, 2000), while probabilistic approaches use ground motion parameters, would be the arias or PGA intensity, for different return periods specified by seismic risk. Among the many proposed refinements of the method (Gallen et al., 2015; Grant et al., 2016; Kaynia et al., 2011; Saade et al., 2016), the most challenging aspect remains the availability of reliable data on the strengths of soils and rocks (Dreyfus et al., 2013) and ground movement. The latter is is is important, because predicting topographical amplification in the steep areas where most landslides occur is the most difficult (Meunier et al., 2008; Sepúlveda et al., 2005). However, the simplified Newmark approach has the advantage of capturing basic physical processes (Allstadt et al., 2017). However, to predict EQTLs, this approach requires assessing voltages and mechanical properties according to topography. Statistical models predict the probability of landslides occurring for a particular area. Most approaches integrate seismic measurements, such as PGA, with field conditions, would be slope steepness, altitude, drainage distance or lithology (Jibson et al., 2000; Lee et al., 2008; Miles & Keefer, 2000, 2001, 2007, 2009). Marc, Hovius and Meunier (2016) and Marc et al. (2017) introduced a model that considers land slippage as a function of the reduced resistance of the substrate caused by soil shaking. The model is based on a set of scaling relationships between the density of spatial landslides, the area of landslide and volume, ground seismic acceleration, the length of the defect, the depth of the earthquake source and the seismic moment. The model is based on the widely observed relationship between landsliding and a minimum level of soil shaking (Harp & Jibson, 2002; Khazai & Sitar, 2004). Soil agitation depends on the source mechanism, wave propagation and attenuation along the surface path, as well as site and targeting effects (Boore & Atkinson, 2008). For reasons of simplicity, attenuation is supposed to be linear, directionality is neglected, and the site's response is taken as a constant average response across the entire landscape. Therefore, this model does not retain any spatial information and predicts only a bulk volume or landslide area above the affected landscape. The model believes that several seismic sources can shake the landscape and that the total volume of landslides reflects the sum of all these point sources. This modeling approach ignores the static variability of the material strength and the possibility that the resistance of the substrate will gradually change during prolonged shaking. The model uses a calibrated relationship between the length and timing of the earthquake rupture (Leonard, 2010), adjusted for the earthquake mechanism (strike-slip or dip-slip). For a magnitude greater than 6.75, saturated acceleration and larger earthquakes do not cause higher landslide densities. The model of Marc, Hovius and Meunier (2016) allows the anticipation of volumes of landslides in bulk, given the basic information about the magnitude of the earthquake, the location of the source, the mechanism and the topographical steepness. These parameters can also be assessed rapidly after an earthquake to obtain a first constraint on coseismic landslides or combined with a distribution of the size of landslides (Hovius et al., 1997; Malamud et al., 2004). Logistical regression (Nowicki et al., 2014; Nowicki Jessee et al., 2018; Parker et al., 2015) and fuzzy logic (Kritikos et al., 2015) are among the most used data-driven methods. Statistical methods also to generate susceptibility maps based on the distribution of 2005 Kashmir coseismic landslides (Kamp et al., 2008), 2008 Wenchuan (Y. Wang, Song, et al., 2016) and 2015 Gorkha (Shrrestha et al., 2018) earthquakes. However, maps cannot be easily used to assess the susceptibility of landslides in future earthquakes, with likely different characteristics and different patterns of landslides. Historical earthquakes can provide useful information if inventories of coseismic landslides are available, but there are very few cases where such inventories are available for earthquakes with different magnitudes in the same area. Little is known about whether and there are models of historical earthquakes and landslides induced by earthquakes (Cauzzi et al., 2014; Nowicki Jessee et al., 2018; Tanyas, Allstadt, & van Westen, 2018) refers to earthquake scenarios in a given area. This would require the generation of a ground motion prediction map for each scenario earthquake. Modelling soil movement for potential earthquakes and integrating results with statistical models trained on historical EQTL inventories could be a future research element, perhaps also as input into probabilistic seismic risk patterns. Earthquakes in mountainous regions can also trigger a small number of very large landslides (>107 m3). Examples of very large EQTLs from 1900 include the following: Osoi, Tajikistan, 2 × 109 m3 in 1907 (Evans et al., 2011); Bairaman, Papua New Guinea, 1.8 × 108 m3 in 1985 (King et al., 1987, 1989); Tsaoling, Taiwan, 1.3 × 108 m3 in 1999 (Chigira et al., 2003; Tang et al., 2009); Hattian Bala, Pakistan, 8.5 × 107 m3 in 2005 (Dunning et al., 2007; Owen et al., 2008); and Daguangbao landslide, China, 109 m3 in 2008 (Huang & Fan, 2013). It is believed that some massive slips of prehistoric terrain were triggered by earthquakes, such as the sliding of the Syseah cliff, the Zagros Mountains, Iran, one of the largest sub-aerial landslides on Earth, with an estimated volume of about 4 × 1010 m3 (Roberts & Evans, 2013). Susceptibility to the generation of very large EQTLs varies depending on the geological and geomorphic setting. Sediment storage landscapes, such as loess land and sensitive clay basins, are particularly sensitive. Many large landslides were triggered in the Loess Plateau in China (Sun et al., 2017; Zhuang et al., 2018) and other loess-mantled parts of Central Asia (Evans et al., 2009; Ishihara et al., 1990) in historical time. 1920 earthquake M 8.5 Haiyuan (Zhang & Wang, 1995; Zhang & Wang, 2007), for example, caused the large-scale collapse of the loess slopes in the Chinese province of Gansu and the emergence of many catastrophic flows, some of which destroyed rivers that destroyed significant lakes. In the Champlain Basin of eastern Canada, february 1663 triggered the large-scale collapse of subaerial and submarine slopes, including massive (>107 m3) sensitive landslides (e.g. Filion et al., 1991; Syvitski & Schafer, 1996). Large EQTLs are of interest from a geomorphic perspective and risk management due to (1) the contribution they make to coseismic denuded and sediment transfer in the affected landscapes, (2) the direct coseismic hazard to the human activities they pose, (3) the postseismic secondary risk they generate through the formation of landslide dams (Evans et al., 2011; Fan, Xu, van Westen, et al., 2017) and the immediate and long-term geomorphic impact of landslides and the failure of the landslide dam and (4) their tendency to persist in the landscape, both as scars and as deposits. Predicting the possible locations and volumes of very large EQTL is not possible with the current state of knowledge. There are no studies that successfully predict the location and extent of very large landslides as failures for the first time. However, forensic studies on the factors controlling their occurrence (Evans et al., 2009) could lead to a better understanding of their causal mechanisms. These, in combination with improved methods of collecting global datasets on geology (structural), geomorphology and topography and improved soil shaking modeling and topographic amplification, could lead to the prediction of potential large landslide sites in the future (Tropeano et al., 2017). Somewhat like very large landslides, few studies have addressed the risk of landslides resulting from earthquake-induced landslides. Since landslide dams are hazards of earthquakes, any hazard and risk assessment should investigate every component of the risk chain: earthquakes → coseismic/post-seismic landslides → dam failure → → dam.break flooding. Event trees are a standard tool for capturing such hazard chains (Gill & Malamud, 2014) by quantifying and multiplying the probabilities of possible mutually exclusive outcomes occurring in order (Marzocchi et al., 2012; see Figure 14; Fan, van Westen, et al., 2012). The use of event trees requires knowledge of possible interactions between all the results of the process. Probabilistic event tree for earthquake-triggered landslide, flood hazard. The six steps (nodes) go from regional to local scale. A branch that ends with Clone is identical to the central branch. For example, at node 1, the other compartments of magnitude are identical to the central branch of 6 ≤ M < 7. At node 2, Si represents the spatial probability (susceptibility) of landslide type i, where av = avalanche; sl = slide; flw = flow; fal = fall, taking into account the specific parameters of the seismic source (e.g. magnitude, type and length of failure and peak ground acceleration). At node 3, means the distribution of the size of the slope frequency of type i, which can be analysed using the methods set out in 2.1.2. The rapid estimation of the extent and distribution of areas affected by landslides immediately after an earthquake is important for coordinating response efforts; highlighting the coordinates of roads, rivers and critical rescue lines blocked by landslides; And And preliminary estimates of losses (Wald et al., 2003; Wegscheider et al., 2013). This assessment is relevant for the days (and possibly weeks) until the first landslide inventory maps become available. Obtaining such data at sufficient reliability takes time, despite many advanced mapping techniques (Martha et al., 2017). For example, the EQTL inventory for the 2015 Gorkha earthquake took six teams of analysts about a month to create one of the fastest rapid response campaigns ever mounted (Kargel et al., 2016; Robinson et al., 2017). For the Wenchuan earthquake, mapping the inventory of landslides took more than 1 year (Gorum et al., 2011; Xu, Xu, Yao, & Dai, 2014). Progress in post-event sensitivity and hazard modelling may be due to the amount of systematic data collection (Figure 15). Prior to the 1994 Northridge earthquake, general statistical relations were concluded between the magnitude of the earthquake and the area affected by landslides or the maximum landslide distance from the epicenter or the rupture area (Keefer, 1984). Jibson and Harp (2012) found that these distances varied according to tectonic settings and therefore magnitude was an inappropriate parameter for characterizing areas affected by landslides. PGA contours provide more conclusive correlations with the magnitude and density of landslides (Jibson & Harp, 2016; Meunier et al., 2007). Wilson and Keefer (1985) initially proposed a minimum threshold of 0.05 g for coseismic lands, based on data from 40 earthquakes collected by Keefer (1984). Jibson and Harp (2016) analyzed six earthquakes with well-mapped landslide data and suggested a PGA range of 0.02–08 g as a minimum ground acceleration capable of triggering landslides; for snow avalanches, this limit is 0,03 g (Podolskiy et al., 2010b). The 1994 Northridge earthquake was the first for which extensive data on the engineering properties of geological units, ground agitation, slope and landslides were available to enable detailed regional analysis (Jibson et al., 1998, 2000). After the earthquake, real-time digital seismic networks were expanded in many areas (Wald et al., 2003), and the U.S. Geological Survey's ShakeMap system (Wald et al., 1999) was developed to provide near-real-time soil movement parameters for significant global earthquakes. These ground movement forecasts are now commonly used in many Assessments of eqtl sensitivity and hazards (Allstadt et al., 2017). Progress in earthquake landslide modelling studies over time. Earthquake Inventories to calculate the cumulative number of inventories are those reported in Tanyas et al. (2017). Vertical lines indicate the time of significant earthquakes for the research of earthquake landslides (red) and products useful for model development (dotted blue line). Sciencebase refers to the release of openly available inventory warehouse (Schmitt et al., 2017). The 2008 Wenchuan earthquake (Figure 15) triggered the highest number of ever recorded ~200,000; (Fan, Domènech, et al., 2018; Tanyas et al., 2017; Xu, Xu, Yao, & Dai, 2014). Using data from this, as well as the 1976 Northridge and Guatemala earthquakes, Godt et al. (2008) developed a global model for the rapid evaluation of EQTL using ShakeMap ground motion parameters and globally available data to estimate critical acceleration of slopes on a regional scale. Without any formal quantitative validation, the authors pointed out that their model provided successful qualitative results. Parker et al. (2015) prepared a statistical model using two earthquakes in New Zealand (1929 Buller, Murchison and Inangahua in 1968). Tested on the 1994 earthquakes in Northridge, Chi-Chi and Wenchuan in 2008, the authors concluded that their model overturned the appearance of landslides. The 2015 earthquake at Gorkha provided the opportunity to test several models in real time. Immediately after the earthquake, three different maps of landslide hazards, based on methods developed by Kritikos et al. (2015), Parker et al. (2015) and Gallen et al. (2015), were posted online. Gallen et al. (2017) conducted a detailed assessment of the statistical landslide prediction models of Kritikos et al. (2015) and Parker et al. (2015), including quantitative validation using the ozone inventory created by Roback et al. (2018) and concluded that the models significantly exceeded the area affected by the landslide due to inaccurate shake estimates provided by ShakeMap. As time evolved after the 2015 Gorkha earthquake, the focus of EQTL studies shifted from modeling methods back to inventory data, as researchers realized that to improve their models they needed better datasets. The number of EQTL inventories has increased rapidly. Tanyas et al. (2017) collected 66 EQTL inventories from around the world and worked with their authors to make them accessible to the public through the U.S. Geological Survey ScienceBase platform developed by Schmitt et al. (2017). This compilation provides a resource for training models. Using 23 of these inventories, Nowicki Jessee et al. (2018) updated the statistical model proposed by Nowicki et al. (2014) for predicting in near-real time the probability of EQTL, by increasing the network resolution to 250 m and using PGV rather than PGA as the primary input of soil movement. Nowicki Jessee et al. (2018) based the 23-invention model to represent a wider range of earthquake magnitudes, climate and tectonic settings. Although the model still predicts the probabilities of landslide (Allstadt et al., 2018), it has been rigorously tested and validated. Tanyas et al. (2019) conducted a similar statistical analysis to assess sensitivity to earthquake-induced landslides worldwide, an approach to the slope unit, which are field partitions assigned to similar hydrological and geomorphological conditions and processes forming natural landscapes. They used a set of 25 earthquake-induced landslide events were classified on the basis of the similarity of causal factors to determine the most relevant training set to make a prediction for a given landslide event. The results show that the classification of landslide events introduces a remarkable improvement in the modeling performance of many events. The classification of existing inventories can be applied in any statistical, global approach to earthquake-induced landslide events. The Us Geological Survey's Prompt Assessment of Global Earthquakes for Response (PAGER) system (Earle et al., 2009) provides rapid estimates of earthquake and death loss, but does not explicitly take into account losses caused by landslides. This long-recognized problem has initiated research to explicitly include ground failure in PAGER loss estimates (Wald, 2013). In one step toward this goal, the U.S. Geological Survey recently launched a new near-real-time earthquake product, Ground Failure, which takes into account both landslides and liquefaction and provides an overall assessment of hazard and population exposure, as well as geospatial hazard maps. The system takes into account global earthquakes, and any models currently implemented must be applicable worldwide. Input data for such models must also be available globally, and models must produce geospatial maps of probability. Currently, three models are implemented for landslides – those presented by Godt et al. (2008), Nowicki Jessee et al. (2018) and Nowicki et al. (2014), the latter serving as the default model for determining the alert level and displaying on interactive web maps. From October 2018, this system provides qualitative descriptors of hazard and loss estimates. Full integration with PAGER requires quantitative estimates, but further research and development is needed to reach this point. Allstadt et al. (2018) used remote detection and field observations from the 2016 Kaikōura (New Zealand) earthquake to assess the three models currently implemented in the product of the US Geological Survey. They examined the model's performance and how prediction maps changed as ShakeMap ground motion estimates evolved over time. For this test case, the authors concluded that (1) any of the three models were useful for the gross prediction of EQTL models, (2) all models outweighed the danger and (3) the temporal change of ShakeMap models that appears as more data on defect geometry, felt ratios, strong soil movement data are incorporated in the hours to the days after the earthquake has a strong positive effect on production Landslide dams are another important process in the waterfall of earthquake danger that requires a quick assessment. This includes assessments of dam stability, potential dam failure mechanisms and modelling of possible exit floods (Figure 16). The longevity of landslide dams varies from minutes to thousands of years, depending on several factors, including the volume of the dam, dam, form and sorting of the locking material, blockage infiltration rates and sediment and water flow rates in the newly formed lake (Costa & Schuster, 1988). Studies on the longevity of landslide dams (Costa & Schuster, 1988, 1991; Ermini & Casagli, 2003; Fan, Tang, et al., 2012; Peng & Zhang, 2012a, 2012b) can be summarized as follows: 85-87% of dams that apparently failed lasted less than 1 year, 40-71% less than a month and 27-34% less than 1 day. The stability of landslide dams depends to a large extent on their internal composition, which can be problematic to discover whether the time between training and failure is short (Costa & Schuster, 1988; Korup, 2005; Korup & Tweed, 2007). Therefore, rapid forecasts of the stability of landslides must remain prone to high uncertainty. Casagli et al. (2003) proposed the use of geomorphic parameters for a rapid assessment of dam stability as a prequel for more accurate and time-consuming analyses of geotechnical stability (Dunning et al., 2005). After the Wenchuan earthquake, Xu et al. (2009) proposed a rapid hazard assessment matrix, taking into account the dam height, composition material and maximum capacity of the lakes that hit the landslides, which were widely applied in the month after the earthquake. Thirty-two of the dams were assessed as having a high risk of dam violations and were artificially breached by the Chinese government. A proposed hazard assessment model and a mitigation workflow for land sliding dams modified by La Fan, Zhan, et al. (2018). The stability of the landslide dam and their violation is a well-studied topic (e.g. Bonelli, 2013, 2018). Studying probable dam desation mechanisms may involve scenarios of over-optation, pipeline, erosion, suffusion, drive and gravitational collapse (Walder & O'Connor, 1997). For the breach mechanisms, such as the pipelines, major numerical progress has been made (e.g. Rodroto et al., 2017), including discrete numerical modeling with hydromechanical coupling (e.g. Tran et al., 2017). Numerous empirical equations express peak discharge at the site of the dam with measurements such as the volume of the lake sequestered, the depth and the surface (Costa & Schuster, 1988; Evans, 1986; see also Peng & Zhang, 2012a, 2012b, for overview). The optimal approach is to apply a physically based model to simulate the dam failure process, predict the resulting flood hydrograph and map the downstream flood wave routing. A commonly used physical model is BREACH, developed by the U.S. National Weather Service (Fread, 1991). Fan, Tang, et al. (2012) applied BREST to predict flood hydrographs for the Tangjiashan field and used the output hydrographs of the hydrodynamic model 1-D-2-D SOBEK to simulate spatial variations in flood parameters, such as arrival time and depth of flooding. Proxies of herodibility of dam materials and the infringement process were also applied (Chang & Zhang, 2010; Chen et al., 2015; Davies et al., 2007; Shi et al., 2015; G. Wang, Furuya, et al., 2016), as well as a geographical information system model (Croissant, Lague, Steer, & Davy, 2017; Mr. Li, Sung, et al., 2011). Disaster management in earthquake-affected mountain areas focuses on the response phase for immediate aid. This is followed by the recovery phase for rehabilitation and reconstruction (UNISDR, 2017). Depending on the magnitude of the earthquake and the resilience of the affected society, recovery may be highly variable (Figure 12) or may even involve the abandonment of the disaster zone (Zobel & Khansa, 2014). In any case, recovery efforts should be informed by earthquake waterfall hazard and risk assessments, which should be updated periodically (van Westen & Greiving, 2017). EQTL inventories are important data in this regard. However, such inventories typically are not detailed enough to provide storm time marks that reactivate coseismic landslides, unless the images have high repeat rates. Obtaining such data from disaster-affected areas in the event of an earthquake requires collaboration between different organisations and, ideally, local communities, in reporting landslide events, for example through web-based tools (Frigerio et al., 2014). Vegetation monitoring is an important component for the analysis of changes in the risk and risk of post-seismic landslide, as vegetation regrowth can stabilize EQTL areas (Yang et al., 2018). Forest cover and agriculture can be both a cause of post-seismic landslides and elements at risk (W. -T. Lin, Lin, & Chou, 2006; Yang et al., 2018). Monitoring of changes in landslide materials in the postseismic landscape informs estimates of sediment cargoes and flooding in rivers. Drilling and geophysical studies (Jongmans & Garambois, 2007), geometric estimates, physically based models, empirical relationships and DEM-de-difference allow the quantification of landslide volumes. The most commonly used method for estimating the volume of landslides is to rely on regression patterns between the area of landslides and volume (Guzzetti et al., 2009). Although generally applicable, this method can lead to large errors for both individual landslides and total inventories (Laren et al., 2010). More accurate results, although for fewer landslides, come from repeated LyDAR surveys corrected for the disturbing effects of vegetation or other surface objects (Chen et al., 2006; Chen et al., 2013; Tseng et al., 2013). In practice, LiDAR dems before the earthquake are rarely available, so other, less detailed alternatives should be considered, e.g. contour lines or photogrammetry (Fan, Xu, & Scaringi, 2018; Fan, Xu, Scaringi, et al., 2017). Predicting dynamics coseismic landslides and deposits introspection yields on how quickly large-scale mass effects we lose mitigate. In the area affected by the Wenchuan earthquake, Fan et al. (2019) created a post-seismic multitemporal database, which includes 527 debris streams in 244 river basins. They also collected precipitation data from 91 rain pledges, which they correlated to analyze the relationship between precipitation and the characteristics of debris flow (release set date: release: and al., 2018). The study showed that the total number of active landslides decreased rapidly over 7 years (from 9,189 in 2008 to 221 in 2015), and post-seismic reactivations occurred randomly in the studied river basin, making them very difficult to predict. They also concluded that, even with such a large data set, it remains difficult to establish empirical precipitation thresholds due to litological, soil and land cover conditions in river basins and the lack of reliably reported events. Critical thresholds may work for warning purposes in individual river basins, but may be inapplicable to others (Huang et al., 2015). Physical-based models and geotechnical investigations are better suited for analyzing post-seismic changes in precipitation thresholds that trigger landslides. Many studies use regional versions of the infinite slope model, ideally along with hydrological models to incorporate groundwater level changes or precipitation infiltration changes. Examples of these models are SHALSTAB (Montgomery & Dietrich, 1994), SINMAP (Pack et al., 1998) and STARWARS + PROBSTAB (van Beek, 2002). By design, the models with infinite slope apply to superficial, translational landslides. S. Zhang et al. (2018) used a Monte Carlo variant of this model to express the uncertainty of the mechanical parameters of the soil in a pattern of precipitation-induced shallow landslides in the Wenchuan earthquake area. More locally, Chen and Zhang (2015) applied a physically based distributed model to predict precipitation-induced shallow slope failures in two-layer soils under different precipitation conditions. While the 2-D and 3-D slope stability models were originally only applicable at the site scale, modern 3-D models, such as Scoops3D (Reid et al., 2015) and r.slope.stability (Mergi et al., 2014) allow large surface simulations and different geometries for the sliding surface. The parameters of these models must take into account the spatial variation of the geotechnical and hydrological parameters (Reid et al., 2015). Empirical runout models, such as Flow-R, combine flow patterns and probability of failure to create runout susceptibility maps (Horton et al., 2013) and are useful on a regional scale. Models runout, ar Flow-2D, RAMMS, MassMove2D (Beguería et al., 2009), Debris Mobility Model (Kwan & Sun, 2006) and AschFlow (Luna et al., 2016) were widely used to assess the hazards and risks of debris flow (e.g. M. Chang et al., 2017), but require input spooling, sediment concentration, initiation and training volumes. In any case, such models may provide information about the relationships between entries (e.g. consequences (e.g. runout; van Asch et al., 2014). Few of the available physically based models combine slope stability, erosion, drive and runout. Thus, recent attention has been turned to integrated models for hydrological and sedimentary processes. Fan, Lehmann, McArdell and Or (2017) have introduced an innovative way to shape the behavior of landslides at capture scale connecting the initiation of precipitation-induced landslides with debris flow patterns. N. Zhang and Matsushima (2016) developed a soil water mixing model, along with an in-depth integrated particle method for debris flows, taking into account sediment drive and precipitation characteristics in the Wenchuan epicentral area. Bout et al. (2018) has developed an integrated model of slope hydrology, slope failure, mass and drain movements (OPENLISEM), which has allowed for better modeling of hazard interactions that may occur during an extreme precipitation event, such as the initiation of landslides, leakage, landslide, flooding and debris flow. The integrated simulation of hydrology, training and the initiation of debris flow was also tried by Chen and Zhang (2015) in the EDDA model, by Hu et al. (2016) and by Lehmann and von Rütte (2017) in the STEP TRAMM model. These models take into account different processes that interact, including precipitation leakage, erosion, initiation of landslides, debris flows and floods, in different scenarios, for example, for changing the availability of sediments or vegetation stands. Many of these models underestimate landslide volumes due to the use of a simplified landslide initiation mechanism. Risk elements can be classified in different ways depending on the expected risk management application (van Westen et al., 2008), for example, for the design of early warning systems, the distribution of the population in space and time is the most relevant. In planning risk reduction measures, a cost-benefit analysis may be required to compare different alternatives and, in this case, it is important to quantify physical objects, such as buildings, in monetary terms. The interaction of the elements exposed to risk and hazard fingerprints (the area affected by a hazard of a certain intensity) defines the exposure and vulnerability of the risk-to-risk elements. Risk components (danger, exposure and vulnerability) change after an earthquake (Figure 12). Severely damaged buildings and other infrastructure could be demolished. Temporary shelters are needed for the homeless, and choosing appropriate sites is an immediate task (Fang et al., 2011). After the response phase, the risk elements change rapidly during the rehabilitation and reconstruction phases (Dalen et al., 2012; Y. Wang, Zou, & Li, 2015). For the mapping and characterisation of these changing elements, tele-detection combined with voluntary geographic information (Schelhorn et al., 2014) may be useful. Voluntaried Geographic Information has increasingly witnessed the reassessment of disaster risk elements through initiatives such as the Humanitarian OpenStreet Map (Geiß & Taubenböck, 2017). is another essential component in estimating losses and risks caused by natural hazards. It remains a multifaceted component and numerous perspectives have been proposed on how to characterise the best vulnerability (Birkmann, 2006; Fuchs, 2009; Shirley et al., 2012). Engineering and natural nature sciences focus on physical vulnerability as the degree of loss for a particular element or set of elements in the area affected by a hazard, expressed on a scale of 0 (without loss) to 1 (total loss) (UNDRO, 1984). Vulnerability can also be expressed as fragility, reflecting the likelihood of reaching a certain state of deterioration for a certain level of intensity. Factors that confuse vulnerability estimates in EQTLs include different types and landslide processes; the difficulty of expressing the intensity of the landslide; limited information on precursor, trigger and control conditions; little accurate or limited observation al.to estimate the intensity of the process (Papathoma-Köhle et al., 2011); and data on reduced damage, in particular in relation to their buildings or lost lives (Fuchs, 2009). Vulnerability curves express the relationship between the intensity of the hazard (e.g. impact pressure) and the fraction of damage mapped directly after disasters, numerically simulated or obtained from expert expertise (Fuchs et al., 2007; Jakob et al., 2012; Zeng et al., 2012). Dynamic debris flow leakage patterns provide results on the depths, speeds and impact pressures of landslides and debris streams and can thus help assess vulnerability. Care should be taken to use derived vulnerability curves for different types of buildings and environments. Physical vulnerability curves for post-seismic debris flows were derived from the Chi-Chi and Wenchuan earthquakes (Lo et al., 2012; Zeng et al., 2014). When compared to European curves, they differ widely, primarily because the buildings in Wenchuan were newly built after the earthquake and were much more resilient than those used for the development of European curves. In some cases, the vulnerability of buildings or roads can be assessed using differential measurements of the interferometric synthetic diaphragm radar, combined with observations on damage (Bianchini et al., 2015; Peduto et al., 2017). In conclusion (see also the summary in Table 5), few studies provide a comprehensive quantitative risk assessment for post-seismic hydrometeorological hazards due to several challenges: Quantifying spatial hazard chains is difficult. The risk assessment of coseismic and postseismic landslides is hampered by complications in reducing the results of the probabilistic seismic risk assessment at local level, where topographic and soil amplification can modulate the distributions of the size of landslides (volume) and therefore the pattern of physical impact. Reactivation of coseismic landslides controls sediment washing in and out of the local drainage network, but few conceptual models currently deals with the sustained dynamics of the slope, which makes this component difficult to predict. More effort should also be made to model the failure surfaces so that the actual volumes can be calculated. Monitoring the change of risk elements, such as buildings, population, infrastructure, land use and land cover, is essential for post-seismic post-seismic risk post-seismic New tools, such as drones and high-resolution satellite images; voluntary geographical information; and crowdsourcing could prove to be valuable ingredients in multihazard risk assessment. Vulnerability assessment also requires more attention, for example, in the form of comparable curves of physical vulnerability and fragility. EQTLs can weaken and erode large amounts of material through landsliding. This sediment is initially deposited on the hills or bottom of the valley, where it is prone to further rework in sedimentary cascade. Local agriculture fills canals, reduces accommodation and alters the likelihood and severity of flooding (Figure 17), which affects resilience and post-seismic recovery. The long-term relevance of these sediments, for example for landscape developments, is discussed in Section 6. This chapter examines how and how time the transport of sediments is being transported and what may be some of the consequences. A growing literature aims to understand the sequence of sedimentary impact of landslides on channel processes (Chen & Petley, 2005; Croissant, Lague, Steer, & Davy, 2017; Davies, 2017; Hancox et al., 2005; Korup, 2004b), in particular where buildings, bridges and other critical infrastructure are in the way of sediment pulses derived from landslides. The work focused on documenting the transfer of sediments from hills to canals (e.g. Fan et al., 2019; Fan, Domènech, et al., 2018; Tang et al., 2016), the pulse of the resulting river sediments and the aggradation of the river canals (Dadson et al., 2004; Pears & Watson, 1983; J. Wang, Jin, et al., 2016; J. Wang, Jin, et al., 2015; Yaunites et al., 2010), as well as the long-term record kept in the definitions and fan sediments (Davies & Korup, 2007; Schwanghart et al., 2016). Illustration of how sediments from earthquake landslides delivered to the river system can lead to the aggradation of river canals. This leads to a decrease in accommodation and thus increases the likelihood and severity of overbank flooding. The sequence of events described represents a cascade of sedimentary hazards that can prevent reconstruction from earthquakes. (a) River channels prior to delivery of sedimentfrom landslides. (b) Ag-graded channels that can no longer accommodate storm flow, leading to flooding (after Dingle et al., 2017). (c) Images and photographs of the wenjia coseismic landslide and the Mianyuan River in Sichuan, China, immediately after and then years after the earthquake. Landslide deposits were remobilized and developed in a catastrophic stream of debris during the storm on August 13, 2010. Debris flows partially into the Mianyuan River, causing the newly rebuilt Qingping into flooding. The buildings with a red outline were partially buried by of post-seismic debris. The sedimentary response to large earthquakes has been recognized for several decades (Keefer, 1994; Pain & Bowler, 1973; Pears & Watson, 1986) and previous compiled accounts compiled Ambraseys (2009), but the Chi-Chi earthquakes of 1999 and Wenchuan in 2008 stimulated a quantitative understanding of the dynamics of post-seismic sediments. Records of the concentration of suspended sediments and the discharge of water commonly collected at the river fishing stations in Taiwan and Sichuan contributed to the constraint of pre-earthquake conditions and to the quantification of the extent and timing of the disturbance (Dadson et al., 2004; Hovius et al., 2011; Huang & Montgomery, 2012; J. Wang, Jin, et al., 2016). Suspended sediment flows increased as a result of both earthquakes as a result of landslide inflows (Figure 18), largely due to intense rainfall and high spills in subsequent years; this process highlights the role of river transport capacity in regulating the rate of escape of landslides. Records of multi-year suspension landslides and landslides based on events since the Chi-Chi earthquake in Taiwan (a) and the Wenchuan earthquake, China (b). (a) is modified after Hovius et al. (2011), where K is defined as the concentration of normalised suspended sediments for river discharge. (b) shows the volume of landslides before and after the Wenchuan earthquake in the multitemporal inventory of landslides generated by Fan et al. (2019) and downstream data on suspended sediments from J. Wang, Jin, et al. (2015), reflecting the increase in sediments along the Fujiang River affected by EqTL. In both cases, notable seismic improvement of sediment flows gradually breaks down at reference levels over ~5 years, similar to the time scale that post-seismic landslide activity also decreases. The relationship between them remains better understood, but both may have a similar underlying control. Dotted horizontal lines represent the average of the normalized landslide areas (in a) and the volume of landslides in point (b) before the Chi-Chi and Wenchuan earthquake. Vertical dotted blue lines show major precipitation events. Although transport capacity is clearly important for the transport of sediments, in different river basins affected by the Wenchuan earthquake, sediment flows are correlated with the proportion of river basins affected by land slippage more clearly than with hydrological parameters, such as drainage and flow power (Li et al., 2017). Also, in both areas of the earthquake, sediment flows relaxed up to pre-earthquake background levels within a decade (Domènech et al., 2018, 2019; Fan et al., 2019; Fan, Domènech, et al., 2018; Hovius et al., 2011; Huang & Montgomery, 2012), despite the landslide material remaining in the landscape (Figure 18). These observations suggest that there is a complicated relationship between the supply of sediment and the transport capacity of as a result of large earthquakes; at least in the rivers studied after Chi-Chi and Wenchuan, the river transport capacity did not appear to control the flow of sediments and, on the other hand, the transport of suspended sediments appeared to have been at least partially determined by supply. The interaction between transport and supply limitation (Hovius et al., 2000) was different factors following the two earthquakes. In Taiwan, post-seismic typhoons triggered many new landslides, providing new sediment slot slots while increasing river transport capacity. The landslides triggered by the typhoon have moved closer and closer to the river canals over time, emphasizing their potential role as sources of river sediment (Hovius et al., 2011). Fewer new post-seismic landslides have formed in Wenchuan, but debris flows triggered during post-seismic storms have produced much coseismic sediments in rivers (Domènech et al., 2019; Fan, Domènech, et al., 2018. Fan et al., 2019; Tang et al., 2016; S. Zhang, Zhang, et al., 2016). Little detailed data are available on gaging stations for other earthquakes that trigger landslides to quantify changes in sediment flows in suspension, although other data become available (e.g. 2010 Maule earthquake, Chile; Tolorza et al., 2016). Even without detailed time series, analyses of long-term sediment flow records of active seismicity patterns support the idea that earthquakes are important factors of erosion (Vanmaercke et al., 2014, 2017). Anticipatory planning may open up possibilities for immediate postseismic measurements where there is basic data, but monitoring has stopped in recent decades, as is done in Southern California (Warrick & Mertes, 2009). Proactive initiation of reference monitoring would be prudent in locations highly prone to large earthquakes in the near future, it would be in the South Alps of New Zealand or in cross-sectional ranges in southern California. Most data gaging captures only suspended river loads. Once driven by rivers, some of this suspended material is removed in distal sedimentary basins or offshore, where its sequestration may have geodynamic implications. Thicker bed sediments accumulate in channels and alluvial fans. For example, in the Matiri River in New Zealand, Pears and Watson (1986) estimated the amount of sediment that persists from landslides triggered by the 1929 Murchison earthquake. They suggest that much of the landslide material has remained in streams even after 50 years. A study of the small river basins in Japan by Koi et al. (2008) showed that the yields of contemporary sediments in steep rivers were still largely influenced by the 1923 Kanto earthquake. In this regard, Yanetes et al. (2010, 2011) concluded that segments of Taiwan's Peikang River would likely take hundreds of years to evacuate sediments derived from landslides triggered by the Chi-Chi. Stolle et al. (2019) earthquake report that rivers around Pokhara, Nepal's second largest city, are still adapting today to massive agricultural most likely during several powerful medieval earthquakes. Most of the network of active channels is currently incised into thick valley fills that are less than a thousand years old. Repeated studies in the earthquake area of Wenchuan have shown a progressive movement of landslide materials from the hills in the canals between and 2013, representing <10% of the total volume of landslides in 5 years (S. Zhang, Zhang, et al., 2016). In some areas, rivers incised by tens of meters in the Sichuan basin downstream of the source regions of landslide (N. Fan, Nie, et al., 2016); this observation is similar to that reported by Yaunites et al. (2010) for the Peikang River after the Chi-Chi earthquake. The concentrations of the 10Be cosmogenic nuclide in river current grains provide an additional insight into the dynamics of the bed load. This cosmogenic isotope is produced by the interaction of cosmic rays with minerals from several higher meters of the Earth's crust. Landslides remove material from deep with low or negligible concentrations of 10Be (Brown et al., 1995). Since this material mixes with pre-earthquake sediments in rivers, the concentration of 10Be in the current, a weather-resistant mineral, can follow the addition and eventual removal of the landslide material (Wang et al., 2017; West et al., 2014). This approach requires 10Be data before a major earthquake and is limited to the dimensions of cereals that can be sampled and analysed (and for which there are data before the earthquake), but its application in tandem with other studies on the transport of post-landslide sediments is promising. Another promising technical development is the application of seismic techniques to monitor active bed load transport (e.g. Chao et al., 2015; Schmandt et al., 2013); such techniques could find a compelling application in understanding the transport of sediments derived from landslides. Future efforts to quantify sediment response to earthquakes could use more high-resolution satellite images and possibly search large areas with UAVs. For example, Yanetes et al. (2011) used satellite imagery and field studies to provide a regional assessment of the dynamics of multi-year aggradation in central Taiwan to test the influence of intermittent sedimentation on long-term incision rates. However, the removal of bed sediments derived from landslides could take several decades and thus extend beyond most current monitoring programs. Several possible paths are opened to extend this observation window to target sedimentary sinks, including reservoirs, debris basins and dammed lakes as records of post-seismic sediment flows (e.g. Foster, 2010; Lavé & Burbank, 2004). Prehistoric sedimentary archives provide additional information about the sedimentary legacy of earthquakes in the past. These records do not usually have the chronological resolution to quantify the dynamics of sediment transport in but they can provide an insight into the integrated sedimentary response over the decades to millennia (Howarth et al., 2016). The lake sediment can be a useful archive of the entry of post-seismic sediments (Howarth et al., 2016), although the lake turbidites and deformed layers (Karlin & Abella, 1992; Moernaut et al., 2014; Sims, 1973) may not always represent entrances to the surrounding river basins. Along the Anatolian fault line in Turkey, Avsar et al. (2014, 2016) used the geochemical footprint of ultramafic rocks associated with EQTL to fix the timing and duration of sediment pulses, which they found to have lasted 5-10 years. Alluvial deposits can also experience sediment disturbances after earthquakes, a notable example being the alluvial filling around Pokhara, now recognized as the legacy of catastrophic sedimentation following medieval earthquakes (Schwanhart et al., 2016). To date, work in these alluvial environments has focused primarily on identifying characteristic signatures that can be fixed on earthquake events, although Stolle et al. (2019) has rebuilt the former geomorphic surfaces of these valley fillings to deduce high rates of erosion and sediment yields on average over more than half a millennium. Finally, studies of canals and fields that were formed in response to large, non-seismic land-slip entries provide a basis for comparison and thus for estimating the seismic inheritance of the massive gradation of the valley (Korup, 2004b). A few detailed post-seismic sediment budgets highlight that the research still misses a predictive framework for estimating the effects of single earthquakes on the river system, much less a sequence of earthquakes. Predicting the river response to the earthquake-triggered landslide depends on the prediction of the speed at which sediment makes its way from hill deposits into and through the river system over the years to millennia (Figure 19). Schematic diagram illustrating the components needed to construct a comprehensive post-seismic sediment transport model, including key interconnections (red arrows) leading to dynamic changes in connectivity, grain size and flow conditions during the post-seismic evolution of the sediment system. Components are classified on the basis of the assessment of the current relative state of knowledge for each component (ranked out of 5, 5/5 being the best known). For example, landslide mapping is now relatively well developed and sophisticated, and this information can be obtained relatively quickly from satellite images, while information on grain size is limited. Green forms refer to the supply of materials; light purple shapes refers to the transport of materials. Several studies have modeled sediment pulses derived from landslides or debris in the river system as stochastic processes (e.g. Benda & Dunne, 1997; Cui, Parker, Lisle, et al., 2003; Cui, Parker, Pizzuto, & Lisle, 2003), which introduces a probabilistic element to sediment transport patterns following earthquakes (Croissant, Lague, Davy, et al., 2017; Croissant, Lague, Steer, & Davy, However, shaping the sediment dynamics associated with the entry from several thousand or even tens of thousands of discrete landslide sources highlights at least two important questions: What determines the amount of landslide sedimentdelivered to high-order river channels, as opposed to the isolation on the What determines the ability of landslide sediments to move through a river network? Addressing these questions with more field data will help to quantify links in the sedimentary chain and inform predictive modelling. Efforts to address these questions are taken into account in the following two sections. Connectivity refers to both the fraction and the rate of coseismic material that is delivered to the channels, as well as the underlying processes. Landslide deposits not connected to the river system may remain abandoned for long periods of time; such deposits are phased out only by slow, diffuse hillslope transport, would be ground creep. Several approaches can be used to measure connectivity in landslide inventories. Relationships between slope areas derived from digital elevation data can be used to define a threshold catchment area for river channels (Montgomery, 2001), and landslides can then be defined as being connected to channels if their lowest elevation point reaches or intersects the channel (Dadson et al., 2004; Li et al., 2016). However, there are several uncertainties in the application of this method. One is that it is not clear whether rivers in a region affected by landslides have a common threshold of initiation. Another aspect is that the threshold area defining river channels is rarely distinct in the slope space (Montgomery, 2001), partly due to methodological uncertainties (Clubb et al., 2014; Passalacqua et al., 2010) and partly due to the destruction of the canal heads by earthquake-induced landslides. For the inventory of landslides in Gorkha, the uncertainty at the threshold of the canal area resulted in variations of 33-43% of the area of landslides and 49-60% of the volume calculated as riverly connected (Roback et al., 2018). Another challenge is that data resolution can prevent accurate estimates; G. Li et al. (2016) argued that the resolution of digital elevation data can cause ~10% errors in the estimated connectivity. Finally, the physical contact of a landslide deposit with a river channel may not reveal much about sediment transfer rates; become more important in changing connectivity in low-order channels and hill slopes. An alternative measurement of connectivity is the normalized distance from each landslide to ridges and valleys (Huang & Montgomery, 2012; Meunier et al., 2008), so that low values up to an arbitrary threshold indicate a higher probability of material landslide entering the river channels. This measurement may produce results similar to those obtained using surface thresholds, but is also uncertain. However, another alternative is to assess connectivity from the contrast of vegetation, landslides and river canals in high-resolution images (West et al., 2011). Regardless of the method, most studies found that between 5% and 20% of coseismic landslides intersected with rivers (Dadson et al., 2004, for Chi-Chi; Li et al., 2016, for Wenchuan; Roback et al., 2018, for Gorkha). This percentage increases to 40-60% if account is taken of the at least in the wenchuan and Gorkha cases, as larger landslides with a longer drainage duration are more likely to reach the canals (Li et al., 2016). However, this leaves about half the volume of apparently failed landslide in the drainage network. One reason is that seismically triggered landslides are most likely to initiate near ridges (Densmore & Hovius, 2000; Harp et al., 1981; Meunier et al., 2008); they may be less likely to reach the canals than landslides triggered by precipitation. Studies comparing different river basins affected by the Wenchuan earthquake have found that mapped connectivity does not significantly improve the correlation between river basins and post-seismic suspended sediment flows (Li et al., 2016), much less the concentrations of 10Be in river sediments (Wang et al., 2017). It seems that connectivity might not fully reflect the ability of debris to slide through the ground to reach the canals (Figure 20), probably because of one or more of the following: loose and empty material in landslide deposits on the hills could be very mobile during storms, while coarsely clasted could armor landslide debris. Low control channels may not have the transport capacity to move landslide materials, even when they appear to be connected to the river system. Debris flows could cause large fractions of material to slide into the hills and deliver them to rivers. Only high and very rare flood stages could result in landslide materials affecting the canals. Connectivity between landslide deposits and canals modulates the supply of landslide sediments of hydrographic systems. (a) Only the te of large landslide deposits connected to channels is immediately accessible for training; (b) clasted coarse armor debris landslide against erosion; (c) the gradual removal of fine material, leaving behind coarse material which is more difficult to remove; (d and f) river-isolated landslide materials may become mobile through hill transport processes, improved by rolling and enduring landslide deposits; and (e) debris flows can provide a dynamic mechanism for transporting materials from the collusive parts of the landscape into rivers. The coarse material in the head waters can be transported to higher-order rivers through scrap streams or can remain in the place where it sits and abrasions, gradually reducing the size of the grain until it can be transported through floods. Meanwhile, landslide deposits in larger downstream rivers may gradually be reduced as fines stinging, leaving behind boulder lags. Thus, both grain size and connectivity are dynamic properties that change as the sedimentary system responds to landslide inflows. These factors may need more attention in the activity of quantifying connectivity on landslide maps, and a more dynamic view of connectivity could help reconcile the seemingly juxtaposed limitations of transport and supply of postseismic sediment flows. Further progress could be made through detailed analyses of the teledetection and on-the-spot surveys of and the arming of landslide deposits. For example, studies monitoring erosion activity, such as rilling and gullying of landslide surfaces indicate that more than 6 years after the Wenchuan earthquake, the rate of this erosion has decreased over time, without any clear pattern (Domènech et al., 2019; Fan, Domènech, et al., 2018; Tang et al., 2016). The size of the cereals determines whether and how often landslide sediments delivered to river channels can be transported. If the flow conditions exceed the critical stress of shearing the bed load (Shields stress) and initiate the movement of particles depends on the depth of the local flow and leakage. Smaller grains have less critical stress and require a lower flow depth for transport, while larger clasts move only in large and rare streams. Few studies have quantified grain-sized distributions of landslide deposits and fewer still link them to river sediment dynamics. Land sliding inflows into rivers may increase the size of load cereals in a measurable way (Attal et al., 2015; Attal & Lavé, 2006). Several studies have attempted to physically shape the links between the size of coseismic deposits cereals (e.g. Qi et al., 2012; Wang et al., 2013; L. M. Zhang, Xu, et al., 2011; S. Zhang, Zhang, & Chen, 2014) and sediment volumes delivered to rivers, but accurate measurement of grain-sized distribution of landslide deposits is difficult (Casagli et al., 2003). Therefore, a more systematic understanding of the effects of EQTLs on grain-sized river distributions is important for a full understanding of the dynamics of post-seismic sediments. Understanding the catastrophic degradation of rivers fuelled by debris from coseismic landslides has become more diversified due to detailed records of water and sediment discharge before and after chi-chi earthquakes in 1999 and 2008 (after the above detailed). However, the response of alluvial systems to individual earthquakes or sequences of earthquakes still rarely occurs in long-term studies, especially those focused on sedimentary deposits. How long do the terraces and fan depots that formed in the years after the strong earthquakes on the valley floors persist? Studies, in general, devoted to such forms of relief rarely invoke earthquakes and routinely resort to long-term tectonic changes in the baseline, climate oscillations or human disturbances. For example, terraces with 100-200 m-thick fans intertwine many rivers in the Central Taiwan Zone (Hsieh & Chyi, 2010). The radiocarbon ages of these river deposits and debris flows are widely scattered in the holocen with no clear link to major climate change, so episodic disturbances earthquakes and tropical cyclones seem more plausible generators of such thick and dissected valley fills. Some large alluvial fans could have formed or at least changed catastrophically during or shortly after strong earthquakes and could thus contain valuable information about the timing and extent of the sedimentary impulses that followed (M. -L. Hsieh, Liew, & Chen, 2011; Keefer, 1999; Langridge and 2012). An example is Pokhara, which appears to be built entirely on a 120-km2 fan-shaped warehouse in the Upper Himalayan Annapurna Massif. Detailed topographical analysis, radiocarbon dating and sediment provenance work indicate that 4-5 km3 of upper Himalayan sediments were catastrophically stored up to 70 km from a small source area of more than 3,500 m a.s.l. (above sea level) in at least three major pulses that coincide with documented data of strong earthquakes from the medieval era (Schhart et al. , 2016). These episodes of rapid valley degradation filled nearly a dozen tributary mouths for upstream distances of up to 8 km and have several lakes that persist. Such settings provide much-needed insights into the response and recovery of rivers from EQTLs over decades to centuries, if not millennia. In coastal areas, earthquake-induced sediment pulses can leave long-lasting signatures in the form of large dune ridges that form at river mouths in response to massive amounts of river sediment transferred through the coastal drift. Parallel dune ridge sets, in particular, may contain sedimentary signals of past earthquakes (Goff et al., 2008) and represent the coastal member of a suite of relief forms in the seismic scale of cascading processes. This conceptual model invokes EQTLs which increase the loads of river sediments, which in turn promote the construction of coastal dunes, given the sufficiently short transport distances. The model also predicts that more dune ridges along active tectonic coasts would not only record sediment flow driven by river discharge, sea level and climate change, but would also carry a fingerprint of past earthquakes. From these observations, we infer that a global question on the dynamics of postseismic sediments is the length of time during which the landslide material remains in the landscape. The time required to dispose of the landslide material also determines the duration of cascading hazards. The time of stay and washing of landslide debris is very variable. Reported post-seismic suspended sediment pulses are mostly short-lived (<10 years), but rough sediment degradation in river channels takes longer, and many of the largest landslide deposits remain visible for millennia at least. The general understanding of what these time scales control depends on building a more complete picture of the sediment cascade from hill to channel (Figure 19). In view of this objective of developing a quantitative model for sediment dynamics following large earthquakes, we identify several directions for future research: Collecting more data from future earthquakes proactive monitoring of sediment flows in seismically active regions and preparation for events where baseline data already exist. Development of other high-resolution image methods for tracking post-seismic sediment pulses and channel changes, for example, through the use of high-resolution EMD. Use of lakes, reservoirs and debris basins as post-seismic sediment recorders Quantify connectivity as a dynamic property and build processes-based models of complete sediment cascade from hill slopes to canals. Development of techniques for effectively measuring the size of grain of landslide materials and its modifications over time. Landslides that heal hills and deposits that block river valleys are the most obvious geomorphic signatures of earthquakes in the mountains. However, the impact of EQTLs may persist long after the scars have been revegetated, the deposits eroded, and the river canals discharged. The increase in data on the extent and distribution of EQTLs has contributed to our understanding of the long-term effects of landslides as erosion agents. One question in this regard is that of a balance of seismic mass: Is the net effect of large earthquakes to add material that builds topography or to remove material by landslides? Quantitative evaluations of the EQTLs legacy were largely based on statistical models, and more recent work added numerical simulations. The statistical approach uses EQTL size distributions to deduce average regional erosion rates (Keefer, 1994). These density-adjusted rates can be compared with tectonic lift rates (G. Li, West, et al., 2014; Marc, Hovius, & Meunier, 2016). This strategy depends on appropriate statistical models, sufficient sample size, negligible mapping bias and robust error propagation when combining earthquake return times with the number of landslides. Long-term projections should take into account uncertainties related to the frequency and extent of earthquakes triggering landslides, how landslides respond to a certain intensity of shaking, the effect of precursor triggers and slope instability, and the gradual removal of landslide materials. Numerical simulations are based on some of these assumptions, but also include physical constraints on slope stability and transport laws for river processes (Croissant, Lague, Davy, et al., 2017). The contribution of large individual earthquakes to the rise and erosion of mountain belts was addressed with increasing detail after the Chi-Chi earthquakes of 1999 and wenchuane in 2008 (Hovius et al., 2011; Parker et al., 2011), after Keefer (1984, 1994, 2002) and Malamud et al. (2004). These studies have stimulated a greater interest in the concept of a rate of seismic erosion that refers to the rate of erosion of medium coseismic land landslides over time (G. Li et al., 2017). Earthquakes also contribute to weathering by physically destroying rock masses in mobile regolith (Clarke & Burbank, 2010, 2011) and the commination of rock materials detached in coseismic landslide. Quantification of the contribution of these processes is open to research, although this research is challenged by precise measurement of soil and regolit depths in steep topography and also by the development of a theory of the coseismic regolith generation. Landslide volumes provide a prime estimate of erosion rates for individual earthquakes (Keefer, 1994; Malamud et al., 2004; Marc, Hovius, & Meunier, 2016; Parker et al., al., Direct measurement of the volume of thousands of landslides generated during large earthquakes is difficult to resolve, and most volumetric estimates are based on volume scaling (Lelsen et al., 2010). Long-term landslide rates can be estimated by combining earthquake return periods and associated landslide volumes (Keefer, 1994; Malamud et al., 2004), but in general, these relationships have uncertainties of a magnitude, because much of the variability in the volumes of coseismic landslides results from contributions of local abruptity, the magnitude of the earthquake, the seismic moment, hydrology, and the strength of the mass of rocks. This uncertainty underlines the need for landslide initiation models that include a wider range of factors and can be the scale of the orogens. In quantifying seismic erosion rates based on landslide volumes, a key hypothesis is that the landslide material is removed from the mountain belts and that this removal does not influence crustal isostasis. The export of sediments derived from landslides in the decade following the Chi-Chi earthquake reduced the surface rise by up to an average of 0.25 m, more than 30% of the added mass of the coseismic rise (Hovius et al., 2011). In Longmenshan along the eastern edge of the Tibetan Plateau, rates of seismic erosion caused by landslides designed for multiple seismic cycles could have a similar magnitude to the rates of denuded derived from sediment yields, cosmogenic radionuclides and thermochronology (Li et al., 2017). But is the entire mass of the coseismic landslide removed before the next earthquake? In himalaya Nepal, large earthquakes have return periods of hundreds of years or shorter (Avouac, 2007), and not all coseismic landslide materials could be eliminated before the next earthquake. In the return of Longmenshan, periods of large earthquakes are longer, and Parker et al. (2011) support relatively few remnants of relic landslide that precede the wenchuan earthquake. Numerical simulations suggest that rivers have the ability to remove material relatively quickly compared to the return time of such large events (Croissant, Lague, Steer, & Davy, 2017). However, earthquake-induced landslides accounted for 20-80% of the total volume contained in the 10 largest landslides (Marc, Hovius, Meunier, Gorum, & Uchida, 2016; see Figure 21 to illustrate the importance of the 10 largest landslides in erosion volume for the Wenchuan earthquake). These largest landslides typically exceed 0.1 km3, and landslides of this size have persisted for thousands of years in mountain belts around the world (Korup et al., 2007). The rate at which sediments generated by landslides coseismiccoming from the mountain ranges could largely depend on the rate of erosion of large landslides. A compilation of the fate of large landslide deposits in the mountains worldwide residence documents of up to tens of millennia; some potential EQTL deposits in mountain forellands are several million years old. Specific yields of sediments from the erosion of large landslide bodies may exceed 105 105 in the first year after the location and remain elevated to >103 t•km–2•year–1 thousand years later (Korup, 2012; Stolle et al., 2019). The persistence of these large deposits presents challenges in estimating long-term seismic erosion rates only from landslide volumes. Summary of the concept of mass earthquake balance using the Wenchuan earthquake as an example. (a) Cartoons illustrating the mass balance between the rise of earthquake-induced rocks and the river export of sediments facilitating landslides; (b) the diagram showing the calculation of the volume balance for compression earthquakes in the context of an elastic semi-space model (modified after Marc, Hovius, Meunier, Gorum, & Uchida, 2016); (c–f) for the Wenchuan earthquake, coseismic displacement string profiles, landslide volume and changes in net volume calculated in 1-km-width bands (difference between coseismic displacement and landslide volume, taking into account the entire volume of landslides and the volume of landslides, excluding 10 landslides), projected in perpendicular directions and parallel to fault marks (data on coseismic displacement in Fielding et al. , 2013; data on landslides at G. Li , West, et al., 2014); (g) the sight of the Wenchuan earthquake mapping, landslides, fault marks and swat profile in the context of the Tibetan Plateau and the Longmenshan Mountains. Yellow dots are coseismic landslides. Most studies agree that the total volume of coseismic landslides during a strong earthquake is substantial compared to available data on total erosion flows (Dadson et al., 2004; Hovius et al., 2011; Li et al., 2016). This observation has led to the question of whether earthquakes in compression media mainly generate topography by shortening and thickening crustals or whether they mainly destroy topography by erosion (Figure 21a). Some studies have balanced the relative magnitude of rise and erosion as a result of multiple earthquakes by using semispatial elastic models to simulate lifting (Figure 21b) and have suggested that some magnitudes of earthquakes (M > 8) may be more important than others for the net topographic increase in mountain belts (G. Li, West, et al., 2014; Marc, Hovius, Meunier, Gorum, & Uchida, 2016). But reconciling the combined effects of lifting, weakening rocks, and erosion of earthquakes in relation to aseismic rise, chemical weather, and river incision might merit more consideration in the future. Pushing and reverse failure accommodate a certain fraction of the shortening associated with the compression mountain belts. However, estimates for the total amount of permanent vertical deformation caused by these earthquakes differ in a wide sense (Avouac, 2007; Meade, 2010). Models considers that the role of earthquakes in the generation of topography is minor, as the associated movement acts in opposition to the accumulated interseismic strain (Simpson, 2015). For example, geodesic data indicate that the high Himalayas have decreased by 0.6 m above the plan for the failure of the 2015 earthquake at Gorkha Gorkha and al., 2016). However, the permanent deformation associated with earthquakes is recorded by the lifting of the geomorphic surfaces (Kelsey, 1990). This permanent deformation reflects a combination of inelastic strain that accumulates in the upper crust (Baker et al., 2013), redistribution of mass by gravitational and viscous deformation (King et al., 1988; Simpson, 2015) and mass redistribution through erosion and deposition (Densmore et al., 2012; King et al., 1988; Molnar, 2012), as well as other aseismic processes in the lithosphere, would be the crustal flow (Royden et al., 1997). The relative contributions of aseismic and coseismic processes to surface rise remain controversial. For mountains that multiply on the Tibetan Plateau, plausible models of final members support either an entirely aseismic origin (Royden et al., 1997) or an entirely coseismic origin (Hubbard & Shaw, 2009; Wang et al., 2012). What is the role of earthquakes in the evolution of terrestrial forms if aseismic processes dominate the rise of rocks? This open question might be important to understand the geodynamic consequences of seismically triggered landsliding. The question of whether and to what extent EQTLs contribute to the balance of mass or volume of seismic deformation remains open. Even if seismic erosion does not balance seismic elevation, landslides can have important effects on the long-term evolution of orogenic systems. Landslide-based seismic erosion can help to focus denude near active defects (Li et al., 2017). The resulting discharge could, in turn, affect seismicity (Steer et al., 2014) and generate possible feedback between earthquakes and erosion. EQTLs can also change the morphology of hillslope, as is evident from pre- and post-seismic elevation patterns. For example, Ren et al. (2017) reported a decrease in the average angle of the longmenshan hill after the Wenchuan earthquake. Landslides can also affect topographical development by reorganizing the drainage basin. Very large landslides can establish the spatial pattern of river incision in high-pitching landscapes, such as Papua New Guinea (Hovius et al., 1998). On a smaller scale, Dahlquist et al. (2018) found that a subset of landslides triggered during the Wenchuan and Gorkha earthquakes, as well as during Taiwan's Typhoon Morakot, caused the division by cutting ridges. By changing the distribution of river erosive power, these migrations could generate feedback between seismicity and land slippage. Thus, including the spatial distribution of landslides, could be important for the development of landscape evolution models. In short, earthquakes are important for the erosion of steep mountains, but we do not yet have a systematic understanding of how they contribute to the global erosion budget. Currently, the lifting of and coseismic erosion are treated independently of each other and without the context of interseismic erosion processes; this separation limits our ability to assess the role of large earthquakes in the evolution of Landsliding coupling patterns, hillslope erosion, and fault dynamics could show a way to fill this gap. geomorphic impact of EQTLs over decades, centuries and millennia? What evidence remains in the landscape that could help diagnose the timing and intensity of earthquakes in the past? Few studies cover the topographic fingerprints of EQTLs beyond time intervals of several years or decades, although studies indicating that human memory is short in this regard. For example, much of the extensive deposits discarded by the catastrophic rock/ice avalanche and debris flow from Nevados Huascaran, Peru, mobilized during a strong earthquake in 1970 (Evans et al., 2009) now support new settlements and agricultural fields. By deciphering the long-term relevance of EQTLs, we try to identify clear geomorphic legacies (Table 6). Despite detailed information on EQTLs models and characteristics, we still face substantial deficiencies when it comes to reversing this knowledge and determining a seismic trigger from the evidence given on landslides (Jibson, 1996, 2009). In the steep mountains, EQTL that forms close to the ridges could be distinguished from landslides triggered by precipitation, which tend to initiate in areas of topographical convergence near the bottom of the valley (Densmore & Hovius, 2000; Hovius et al., 2011; Meunier et al., 2008). In these cases, the eQTL-dominated landscapes may have more hovering hills with fewer inner keys (Densmore & Hovius, 2000), although these features have been challenging to identify themselves on the ground (Clarke & Burbank, 2010). The researchers speculated that unique deposits or especially grouped giant landslides (≫106 m3) and also long-lasting proxies of strong soil shaking in the past are possible (Crozier et al., 1995). However, without a visible statistical difference between precipitation and earthquake-induced landslide distributions (Malamud et al., 2004), separating the effects of extreme earthquakes from those of extreme storms will be difficult. For a given region and denude rate, the largest landslide deposits should be most likely to remain for the longest period (Korup & Clague, 2009), and some of the largest still recognized landslide deposits have survived in semi-arid landscapes towards arid landscapes for several million years, would be in the area of active subduction in northern Chile (Mather et al., 2014). Table 6. Summary of topographical legacies of earthquake landslides Possible effects of earthquake landslides on large-scale topography • Stimulation of erosion losses in seismically active mountains; for example, the determination of the mass or volume balance (Figure 21; Hovius et al., 2011; G. Li, West, et al., 2014; Marc, Hovius, Meunier, Gorum, & Uchida, 2016 Parker et al., 2011) • Denotation of focus along seismically active mountain fronts (Li et al., 2017) reorganisation of drainage basins (Dahlquist et al., 2018; Hovius et al., 1998) Potential landslide prints and a few studies for example • Gliding landslides associated with earthquake-induced landslides in landscapes with fewer inner keys due to preferential initiation near ridges (Densmore & Hovius, 2000) • Groups of huge landslides, volume ≫106 m3 (Crozier et al., 1995), with mechanistic analysis of the back to distinguish seismic trigger (Jibson , 2009; Jibson & Keefer, 1993) • Recectives of landslide lakes (Korup & Wang, 2015) • Bowl-shaped source areas (Turnbull & Davies, 2006) • Alluvial deposits from landslides (Schwanhart et al., 2016) • Parallel ridges of dunes along the coasts (Goff et al., 2008) Smaller giant land slides are divided, drainage is determined , disruptors or dam rivers, create long-lived channel knicons or even cause drainage overturns; the largest defects reported to the rock slope buried hundreds of square kilometers under thick sheets of hummocky debris (Shoaei, 2014). Even where landslide lakes do not survive, their former reservoirs usually retain the edges of lake sediments and delta deposits. Groups of huge landslide deposits are not limited to high topographical terrain, but can also affect indistinct, low-gradient, semi-desert landscapes, such as the Caspian Sea basin, possibly providing evidence of rare but powerful intraplate earthquakes (Pánek et al., 2016). The spatial grouping of landslide lakes, even if equal in age, is an ambiguous indicator for the inferalping of earthquake intensity isoseismals, as strong storms can form numerous lakes more or less coevilly (Korup & Wang, 2015). However, even a clear geomorphic footprint of large-scale slope instability may be difficult to interpret. Turnbull and Davies (2006) proposed that the distinct source bol-shaped areas of many deep rock slides, triggered by the earthquake, should be very similar to glacial circuses. The implication is that seismic prints on the rocky slopes of glacier terrain could be clearly visible and persistent, but indistinguishable from true glacial circuses. Mechanistic analyses of the back using the balance of force or moment limit may indicate in the first order whether the aseismic causes and only the triggers were sufficient to mobilise a large landslide, thus excluding a possible seismic origin (Jibson, 2009; Jibson & Keefer, 1993). However, the limitation of the required input parameters generally passes with a view to the previous history of disturbances for the hills in question (Parker et al., 2015) and how close it was to failure before the trigger earthquake. Systematic work on faults of subaque slopes, including turbidites, mass transport complexes and debrilates in seismically active regions, are largely based on the interpretation of earthquake triggers, with great potential to document the longer-term geomorphic impact, in particular when combined with independent land records (Beck, 2009; Leihold et al., 2018; Van Daele et al., 2013). Large-scale land slippage can alter biogeochemical flows with consequences. In particular, landsliding can alter the carbon flows associated with both chemical weather and eroded organic material in the hills (Emberson et al., 2016, 2017; Frith et al., 2018; Hilton, Galy, et al., 2011; Hilton , Meunier, et al., 2011). Although changes in these flows are small in relation to the annual flows associated with photosynthesis, breathing and consumption of fossil fuels, when integrated over 105-106 years of geological time, even small disturbances in weather rates and the burial or oxidation of organic carbon can have dramatic effects on the concentration of CO2 in the atmosphere (e.g. Caldira et al. , 1999). The assessment of these possible links requires the quantification of the effects of individual events, and this quantification is essential to determine whether EQTL acts as the source or sink of CO2 in the atmosphere. Growing datasets, although largely associated with storms, have documented the effect of landslides in the erosion of standing vegetation and in the delivery of organic materials to hydrographic systems where it can be easily eroded, mainly in the form of organic carbon in particles (POC; Clark et al., 2015; Hilton, Galy, et al., 2011; Hilton, Meunier, et al., 2011; Ramos Scharrón et al., 2012). Rivers that drained Longmenshan also had POC flows twice as high as the Wenchuan earthquake before (J. Wang, Jin, et al., 2016). Estimating the implications of the carbon cycle of these changes depends on knowledge of both the eroded carbon source and its destiny (Figure 22). Organic carbon from landslide material can be derived both from base rock carbon (petrochemical carbon [POCpetro]), which has been sequestered from the atmosphere for millions of years, and from carbon derived from recent photosynthesis biomass (biospheric carbon [POCbio]), which comprises the soil and plant carbon that has lived on the hills and can vary from age to thousands of years (Blair et al. , 2004; Hilton et al., 2008; Hilton, Galy, et al., 2011; Hilton, Meunier, et al., 2011). Once mobilised during the slide, both POCpetro and POCbio can be kept in landslide deposits, removed by rivers and buried in downstream or oxidized sediments during river transport, releasing carbon back into the atmosphere in the form of CO2. If POCbio is kept in sediment while new vegetation recreates on old landslide scars, it will serve as a CO2 sink. If POCbio is oxidized, it will have little effect on long-term carbon flows. Preservation of POCpetro will also have little effect, although POCpetro oxidation will act as a source of CO2 by releasing old carbon into the atmosphere. The net effect of landliding on the organic carbon cycle thus depends on several factors that are a challenge to (Figure 22). Summary of possible effects of earthquake landslides (and landslides in general) on the carbon cycle. The effects of the carbon cycle are related to the modification of organic carbon and inorganic carbon flows, the latter by chemical reactions of weathering. Weather. arrows reflect connections that are not well known; for example, there is clear evidence that biospheric organic carbon is stripped of hills and eroded by landslide deposits, but its final fate (burial vs oxidation during transport) is not well understood; this leaves open questions about the direction in which landslides change atmospheric CO2 emissions. Landslide deposits can also act as weather reactors, reflected in improved solved streams of drainage waters (Emberson et al., 2016). But much of this solute load can be derived from sulphide oxidation and carbonate integrating (Emberson et al., 2017, 2018), at least where they work. Silicate mineral weathers produce alkalinity that consumes CO2 by precipitation of carbonates such as CaCO3 (Berner & Raiswell, 1983; Revelation & Suess, 1957). Instead, sulphide oxidation can release CO2 into the atmosphere, with effects that can last millions of years (Torres et al., 2014). Thus, the net effect of chemical weather in EQTLs on the carbon cycle could depend on the balance of silicate versus dissolution of sulphide, which in turn depends on lithology. Quantification of the carbon cycle footprint of earthquakes should also take into account the hydrological response to seismic shaking (Manga & Wang, 2007). After the Wenchuan earthquake, solute flows increased mainly in Min Jiang, the largest river that drains the epicentral area; however, these increases did not match the spatial model of landsliding, suggesting hydrological control (Jin et al., 2016). Systematic changes in the flow path could have equally important implications of the carbon cycle, especially if they support long-term increases in solute flows. But these changes could be difficult to conclusively distinguish from those associated with weathers landslide. Intriguingly, the hydrological role in weathering could be related to land slipping in interesting ways that operate in several directions. The role of groundwater in determining the site's response to seismic agitation is increasingly recognized (e.g. Cox et al., 2012; Wang et al., 2001). Hydrological conditions can be a key control over coseismic land, but these conditions usually remain unresolved. Relationships between hydrology, coseismic terrain, and weather deserve further investigation. Overall, associated earthquakes and landslides could play a significant role in linking tectonic activity to the global climate, but understanding this role remains difficult given the multiple possible effects at work (Figure 22). Work over the past decade has boosted interest in the long-term consequences of EQTLs, with several provocative assumptions about earthquake mass balance, footprint landslides and the carbon cycle and climate effects. Many open questions and major challenges remain, with several interesting directions for future work, including: Developing a more comprehensive understanding of how EQTL relates to long-term erosion, both in terms of whether and how much sediments are removed from the mountain belts and to what extent the landslide plays a role in the production of regolith in the hills. Scalable models for these processes will be important to answer questions about the long-term geomorphic role of EQTL. Landliding and episodic erosion in geodynamic and landscape evolution patterns, so that the concept of earthquake volume equilibrium balance can be placed in the broader context of seismic and interseismic deformation. Such an effort could take into account how landslides affect the spatial distribution of erosion, including through the reorganisation of drainage. Assessment of whether the distinct characteristics of EQTL can be conclusively differentiated from landslides triggered by precipitation in a way that could be applied with confidence for paleo-analysis. Quantification of the effects of landsliding on organic carbon and chemical weather flows in various settings and developing models that can provide an overview of the role of earthquakes on the global carbon cycle. On a large scale, land slippage in the epicentral areas of large earthquakes is now widely recognized as a pivotal geological process, one that raises basic questions about the production of topography by coseismic lifting, destruction of topography by coseismic erosion and routing of excess of material from mountains into sedimentary sinks. Abrupt, active mountain ranges, such as those of New Zealand, Japan, Taiwan, China, and California, have invited many studies of landsliding and red-red face sediments driven by earthquakes. Several destructive earthquakes over the past two decades have allowed more systematic work on earthquake-induced land-sliding mechanisms, patterns and rates. Technological advances now allow for partial or complete automated mapping of tens to thousands of landslides in short periods of time. Such inventories provide a knowledge base for studying how hills and rivers respond to a particular pattern of seismic agitation and soil deformation, including effects such as the spread and attenuation of seismic waves away from an earthquake source and the geological or topographical effects of the site. These inventories have also strengthened research at the interface between seismology, geomorphology and natural hazards. With now more than 60 inventories available, more general models describing the expected landsliding for a given earthquake are possible, helped by interpretation based on physics and modeling. Growing data sets provide the basis for emerging understanding of EQTLs. Controls on the speed at which the earthquake-generated material is produced include how the earthquake's energy translates into local soil acceleration, as well as the effects of soil resistance and rock mass, disturbance and hydrology. These and other controls are likely to be responsible for the wide spectrum of rates at which mountain landscapes respond to comparable seismic disturbances. Long-term observation of mountain areas affected by earthquakes shows that land slip rates remain high years after an earthquake and that debris landslide moves through a waterfall from the hillslopes to river canals and into sedimentary sinks. This potentially adverse chain of events has been known for some time by geomorphologists, but still expects better integration into natural hazard and risk assessments, which would be better activated by a more quantitative understanding. DEMs comparing pre- to post-seismic terrestrial surfaces promise to quantify seismic risk cascade processes and rates with an accuracy that was not previously possible. Our review of a large body of literature identifies several research challenges: Any useful prediction of earthquake-triggered landsliding requires, in addition to more data on case studies, details of the mechanism of the earthquake source and the propagation of seismic waves and data on topographic, hydrological and biological controls on slope stability. In this context, studying the cumulative effects of earthquakes in order can be instructive. The complete cycle of earthquake disturbance and relaxation needs attention and also a definition of the reference state of the landscape. Establishing long-lasting observatories in which all the main parameters of the landscape that may be affected by an earthquake are measured systematically and repeatedly could be a viable option, although this requires a standardised strategy for optimal investment. The rapid and coordinated scientific response after an earthquake is essential for separating coseismic and post-seismic dynamics and their impact throughout the earthquake-affected area. It is desirable that the installation of strategic equipment comes from comprehensive studies, including the repeated and regional acquisition of high-resolution topographic data and the collection of seismic data to track sediments in the landscape. The removal of coseismic debris remains insufficiently documented, and sediment residence times differ between individual earthquakes. Systematic documentation of post-seismic landslides and transport of river sediments is important to understand sediment cascades that highlight how sedimentary signals move in lowland areas and geological records. The removal of debris is specific to the size of cereals, so ideally future models can predict mass flows from the granulometry produced by EQTL in a given litology. Documented transient increases in landslide rates after earthquakes still require mechanistic explanations, given, for example, the damage caused by the seismic agitation of surface hill materials. Controls on the occurrence of very large landslides remain poorly understood. Because they can be extremely harmful, we need criteria by which we can predict where most occur of land. We need criteria to distinguish between very large landslides caused by earthquakes and non-seismic landslides. This would facilitate the use of large landslide deposits to support paleoseismic records where other methods cannot be applied. The landscape's response to is considered to date mainly as regards dry mass waste, rarely taking into account previous weather conditions, groundwater hydrology, vegetation or weather. In order to limit the importance and functioning of these factors, systematic observations are necessary, in parallel with data on seismic agitation, mass waste and sediment transport. The notion of an earthquake hazard cascade with prolonged sedimentary response and associated hazards should be communicated to professional communities dealing with mitigation, resilience and disaster management. Together with these communities, we need to work to predict dynamic disaster scenarios. From a scientific point of view, the community might want to pursue a comprehensive and generic understanding of earthquake-induced landsliding, which can be used to solidly determine its role in landscape dynamics through patterns and observations. Such a general understanding could allow for a more complete exploration of the impact of earthquakes on all Earth's systems. Beyond science, the prize is a safer society in seismically active mountain regions, with quantitative knowledge of the patterns and probability of hazardous processes and the likely economic and personal risks caused by earthquake-induced landslide crises. Cascade of Geographical Hazards – See Danger Chain. River basin – A hydrological area marked by the formation of lines separating the area between two adjacent river basins. Geographic hazard chain – See hazard chain. Continuum Model – A family of numerical methods that are suitable for simulating landslide runouts. They simulate the mass of the landslide as the sum of one or more continuous domains and generally treat the movement as that of a fluid using equations in fluid mechanics to characterize it. Control factor – A physical characteristic, property, or condition that exerts an influence on a particular process. For example, if this process is the initiation of a landslide, various factors (topographic, seismic, hydrological, geological, geotechnical, climatic and anthropogenic) may promote or prevent the initiation of landslides. The angle of inclination of a slope, the permeability of a soil layer or the shaking of the ground caused by an earthquake are all factors that can control the initiation of landslide. Coseismic landslide – A landslide that is a direct consequence of earthquake-induced soil movement. Critical Shields stress – A constant nondimensional value at which sediment transport ation is assumed to begin, which is also known as bed-shear stress. It was proposed by Shields (1936). Dammed Lake – See Landslide Dam. DDA – Discontinuous deformation analysis is a type of discrete element method that explains the interaction along discontinuities in the fractured and articulated rock. Debris flow – A stream of debris (a mixture of fragments of soil and rock) and water that can travel at high speed along steep slopes and inside canals. Debris flows can increase in volume as they move through the incorporation of additional materials from or the surface of the channel or the sides of the channel. Discrete element method – A family of numerical methods that can be used to simulate landslides. In these methods, soil or rock masses are broken down into distinct elements (e.g. spheres and blocks) that interact with each other. Both 2-D and 3-D models are available. Denude – All natural processes that result in wear or progressive loss of the Earth's surface, including weathering, erosion, mass waste and transport (adapted from AGI, 2005). Digital Elevation Model (DEM) – A three-dimensional representation of a part of the Earth's surface. In general, it is composed of soil elevation values arranged in a raster grid with uniform spacing. Data from a digital elevation can be processed to create layers of data and maps with topographical characteristics, such as slope angle, slope curvature and drainage divisions. Early warning system – See the early warning system of landslides. Risk element – The elements at risk are the population, properties, economic activities, including public services, or any other defined values exposed to hazards in a given area. EQTL – Earthquake-triggered landslide. See Coseismic landslide. Erosion – Natural processes by which earth materials are weakened, dissolved or worn out and transported from one place to another. Erosive processes include running water (including precipitation), waves and currents, moving ice and wind (adapted from AGI, 2005). Escarpment – A long rock or steep slope formed as a result of failure or erosion. Escarpment is also used interchangeably with scarp. Escarpment refers to the edge between two forms of relief, while the scarp is synonymous with a cliff or steep slope. Exposed value – Accounting of assets or residents of a region prone to natural disasters. Exposure – See Exposed Value. Defect – A discontinuity in a rock volume, over which there has been a significant displacement in opposite directions as a result of the movement of rock mass, formed mostly by the action of the tectonic forces of the plates. A sloping fault has a foot wall (below the surface of the defect) and a suspended wall (above the fault surface). Fault rupture – Displacement along an earthquake-induced fault. The fault rupture initiates in the center of an earthquake and then radiates outwards along the surface of the defects. FEM – The finished element method. A family of numerical methods that can be used to simulate some behaviors and processes in soil and rock, such as rainwater infiltration and groundwater infiltration, transport of contaminants, heat transfer, stress distribution and deformations. In these methods, soil or rock masses are represented as continuous idealizations that are divided into elements, explicitly only to their joints. Geohazard – A physical process or phenomenon would be an earthquake or landslide, which can occur with or without warning or trigger and which has the potential to substantially affect the natural and (or) built environment. Geological timescale A chronological dating system that links geological layers (stratigraphy) on time. Geotechnical centrifuge – A centrifuge that uses physical and scaled prototype slope patterns to recreate in situ behavior in a laboratory scale. Geological hazard – See Geohazard. GIS – Geographic Information System. A computer software package designed to analyze, manipulate, store, present, or capture geographic or spatial data. Distribution of grain size – Also known as particle size distribution. A list of mathematical values containing information on the diameters of the individual granular matter and their contributing percentage of mass compared to the total mass of all grains. Gully – V-shaped canal, which is a form of erosion relief created by running water on a hill. Danger – The probability that, in a given area, a geological process (such as a landslide or earthquake) will occur within a certain period of time or under the influence of a specific trigger mechanism. For example, a measure of seismic risk is the likelihood of exceeding a certain level of ground movement in a year. Hazard chain – A sequence of hazards that have a cause-and-effect relationship between them. An example would be an earthquake whose shaking triggers a landslide that then dams a river; this creates a lake that can overtake the landslide dam and cause flooding downstream. Hydraulic conductivity – The ability of any solid or semi-solid matter to transport fluid (e.g. water with dissolved minerals and suspended particles) through pore spaces. Hydraulic head – Height above a data plane (usually groundwater mass) supported by hydraulic pressure at a certain point in an underground water system. Hydraulic gradient – Total hydraulic head change speed per flow distance unit in a given direction. Hydrological forcing – Anological cycles of precipitation, snow, snow melting, groundwater circulation and surface water flows. Intensity – The measured value of a physical parameter directly relevant to the potential for deterioration or loss of life from a possible catastrophic natural phenomenon. For example, PGA is a measure of seismic intensity; cumulative precipitation is a measure of extreme weather intensity. Landslide – A general term that refers to a wide variety of landforms of mass movement and processes involving the transport of the downward slope, under the gravitational influence, of soil and rocky material. This general term thus includes slope movements which do not involve actual slippage, such as debris flows, rock falls and rock landslides (adapted from AGI, 2005; Cruden & Varnes, 1996; Varnes, 1984). Ground sliding dam – A landslide blocking a creek or river of and confiscate an upstream lake (landslide lake). Landslide dams can be transient or permanent and have very different physical characteristics, such as size, shape, composition and permeability. Landslide Lake – See The Landslide Dam. Early landslide warning system – A system with which is designed to issue warnings of imminent harmful landslides before they occur. These components and the objectives of the alert issued may differ depending on the scale at which the system is used. The technical component involves physical or empirical models for assessing the likelihood of landslides in defined warning areas, usually by forecasting and monitoring weather variables, to give generalised warnings to communities and institutions working with hazard reduction measures (on a regional scale) or to provide timely alarms for the temporary evacuation of people from areas where , at certain times, the level of risk to which they are exposed may become intolerably high (on a local scale; Segoni et al., 2017). Land landslide inventory – A database that records the location and other features of landslides that are identified by land investigations, teledetection, historical records or other types of research. Inventories of landslides can be divided according to the type of trigger that caused the landslides (e.g. an earthquake and a storm). Mass waste – A general term for the downward transport of soil and rocky material under the influence of gravity (AGI, 2005). Normal defect – A malfunction that is formed by moving down the suspended wall against the foot wall. Paleoclimate – The climate of the past on the longer scale of geological history studied as a specialized field (paleoclimatology). Paleoseismicity – Earthquakes recorded in geological record, most of them unknown in terms of human descriptions or seismograms. Geological records of past earthquakes may include defective layers of sediment and rocks, liquefied sand injections, landslides, raised or suddenly lowered shores, and tsunami deposits. PGA – Peak ground acceleration. Maximum ground acceleration observed in a record of strong motion induced by an earthquake. PGD – Displacement of the peak soil. Maximum ground displacement observed in a record of strong motion induced by an earthquake. PGV – Maximum ground speed. Maximum ground speed observed in a record of strong motion induced by an earthquake. Porosity – A measure of the relative volume of pores in a solid material. Post-seismic landslide – A landslide that occurs after an earthquake, for which the earthquake acted as a predisposing factor, but not as a direct trigger. Predisposing factor – See Control factor. Precipitation threshold – The smallest precipitation event that can trigger a flow of debris or other landslide in a given area. Regolith – The layer of fragmented rock material and formed or stored after it has been transported, which almost everywhere forms the surface of the land and covers or covers the intact base rock (adapted from AGI, 2005). Resilience – The ability of a community to recover the standard of living they enjoyed the occurrence of a natural calamity. Reverse fault – A defect structure resulting from the upward movement of the suspended wall against the foot wall. Risk – Exposure of people and their built environment to geological hazards. For example, a landslide is a type of geological hazard; lives and property likely to be damaged or destroyed by landslides include the risk of being caused by this danger. Risk assessment – A quantitative examination of the oncoming adverse effects of an anticipated (geological) hazard. Risk mitigation – A set of strategies and countermeasures to reduce the risk to people, infrastructure, ecosystem and so on that derive swells from a hazard or a set of hazards. According to the strategy, this is done by directly reducing the hazard and/or reducing its exposed value and/or vulnerability. Runout - Distance a landslide moves. It can be measured along the path of motion, in a straight line, or in horizontal projection, from the landslide to the farthest point of the lime or as a displacement of the centre of mass. Along with the drop in the height of the landslide mass, it can be used to estimate mobility. Shaking table – An experimental device used to study the response of man-made model slopes and structures or components to seismic shaking or other types of vibration by subjecting them to basic soil movement, usually induced by mechanical actuators. Stress shields – A nondimensional number used to calculate the initiation of sediment movement in a liquid stream. It's a non-dimensionalization of shear stress. Site effect – Amplification or de-amplification of seismic motion (acceleration, speed and displacement) at specific frequencies and in specific directions resulting from the interaction between seismic waves, which reach the site after a path-specific attenuation, and local conditions of the site (e.g., the type of material, layered, topography, etc.). Soil classification – A classification of soils based on the relative content of different particle sizes. See also the distribution of the size of cereals. Sliding fault – A fault structure resulting from horizontal movement in opposite directions on a vertical or steep failure. Susceptibility – The generally stable properties of a natural or built system that make it more or less likely to experience a particular hazard in a particular trigger scenario. For example, susceptibility to landslides is a function of slope properties, such as topography (slope angle, etc.), material strength and geological discontinuities that make a slope more or less susceptible to failure during an earthquake or heavy rains. Trigger – A process or event that initiates another process or event. For example, precipitation or earthquakes can trigger landslides. UAV - Unmanned aerial vehicles. A small aircraft or helicopter (quadcopter, hexacopter, etc.) operated without a pilot on board and remotely controlled from the ground. Smaller UAVs are also called drones. Vulnerability – – the likelihood of seeing a certain state of deterioration or the likelihood of seeing a death if a certain measure of intensity is exceeded. An example is the probability of observing the collapse of a building if a PGA greater than 5 m/s2. River basin – See River basin. 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