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INSTRUMENTAL SHAKING THRESHOLDS FOR SEISMICALLY INDUCED LANDSLIDES AND PRELIMINARY REPORT ON LANDSLIDES TRIGGERED BY THE OCTOBER 17, 1989, LOMA PRIETA, CALIFORNIA EARTHQUAKE

Abstract: HARP E.L., *Instrumental shaking thresholds for seismically induced landslides and preliminary report on landslides triggered by the October 17, 1989, Loma Prieta, California earthquake* (IT ISSN 0391-9838, 1993).

The generation of seismically induced landslide depends on the characteristics of shaking as well as mechanical properties of geologic materials.

A very important parameter in the study of seismically induced landslide is the intensity based on a strong-motion accelerogram: it is defined as Arias intensity and is proportional to the duration of the shaking record as well as the amplitude. Having a theoretical relationship between Arias intensity, magnitude and distance it is possible to predict how for away from the seismic source landslides are likely to occur far a given magnitude earthquake.

Field investigations have established that the threshold level of Arias intensity depends also on site effects, particularly the fracture characteristics of the outcrops present. In the case of the M = 7.1, October 17, 1989, earthquake that struck the San Francisco Bay region, thousands of landslides were triggered over an area of 14,000 km². The most numerous were rock falls, rock slides and soil slides.

Landslides produced a complex pattern of surface fractures, in addition to the ones of tectonic origin related to regional structures. The most common deformational features related to earthquake triggered landslides are extension fractures and scarp; the least common are well-defined folds or fractures along the toes.

KEY WORDS: Earthquake Arias intensity, Landslide, Seismic ground deformation.

Riassunto: HARP E.L., *Soglie di scuotimento strumentali per frane ad innesco sismico e rapporto preliminare sulle frane innescate dal terremoto di Loma Prieta, California, del 17 Ottobre 1989* (IT ISSN 0391-9838, 1993).

La generazione di frane provocate dalla sismicità dipende dalle caratteristiche delle scosse così come dalle proprietà meccaniche del materiale.

Un parametro importante nello studio delle frane ad innesco sismico è l'intensità ottenuta da accelerogrammi *strong motion*: essa è definita come l'intensità Arias ed è proporzionale sia all'ampiezza che alla durata della scossa misurata. Data la relazione teorica tra l'intensità Arias, la magnitudo e la distanza, è possibile predire a quale distanza dalla sorgente sismica sia probabile che si verifichino le frane per una data magnitudo del terremoto.

Le ricerche di campagna hanno dimostrato che il livello di soglia per l'intensità Arias dipende anche dagli effetti del sito, particolarmente dalle caratteristiche delle fratture delle formazioni presenti.

Nel caso del terremoto (di magnitudo 7,1) del 17 Ottobre 1989, che ha colpito San Francisco, migliaia di frane si sono innescate su un'area di 14.000 km². I più numerosi sono stati i crolli di roccia, gli scorrimenti di roccia e gli scorrimenti di suolo.

Le frane hanno prodotto un complesso quadro fessurativo superficiale, che comprende anche le fratture di origine tettonica e legate alle strutture regionali. Le più comuni deformazioni legate all'innesco di frane indotte dai terremoti sono fratture di estensione e scarpate, le meno comuni sono le pieghe o le fratture nella zona di unghia delle frane.

TERMINI CHIAVE: Intensità sismica di Arias, Frana, Deformazioni del suolo da terremoti.

INTRODUCTION

It has long been recognized that landslides triggered by earthquakes often account for the major portion of damage and loss of life in earthquakes rather than damage and destruction resulting directly from the shaking or primary faulting. Examples, such as the giant debris avalanche shaken loose from the slopes of Mt. Huascaran in the M 7.9 Perù earthquake of 1970, the rock slide created by the 1959 Hebgen Lake earthquake in southwestern Montana, or the rock avalanche triggered by the 1984 Nagano Ken Seibu earthquake in Japan, serve to emphasize the fact that just one landslide can dominate the entire earthquake hazard scenario of a region.

INSTRUMENTAL SHAKING INTENSITY THRESHOLDS

Because the generation of seismically induced landslides depends on the characteristics of shaking as well as the mechanical properties of geologic materials, it is critical to understand as much as possible about the variation in ground shaking in an earthquake. The most numerous types of landslides generated in earthquakes, rock falls, rock slides, soil falls, and soil slides are extremely sensitive to shaking levels, and their initiation has been correlated with Modified Mercalli Intensive levels of shaking (KEEFER, 1984; HARP & *alii*, 1981, WILSON & KEEFER, 1985).

Until recently, the data most often available on the patterns of shaking intensity from an earthquake were MMI

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contours prepared from observations of damage to structures and utilities. Even now, in most earthquakes, there is a scarcity of instrumental strong-motion data, especially in the near field. Although MMI data is commonly available from most earthquake investigations, its use in determining the detailed variations of strong shaking from an earthquake is of limited value. MMI values assigned by investigators are qualitative and interpretive, based on observations and are often subjective and observer-dependent.

Where strong-motion data exist, we prefer to use an intensity based on a strong-motion accelerogram. Arias intensity, or I_a, was originally defined by ARIAS (1970). Arias intensity is defined as an integration over time of the acceleration squared from an accelerogram:

$$I_a = \pi / 2g \int_0^{\infty} [a\ (t)]^2\, dt$$

where a is acceleration, t is time, and g is the acceleration of gravity. Arias intensity is expressed in units of velocity, usually m/sec. Because the calculation of Arias intensity involves the entire record, the value is proportional to the duration of the shaking record as well as the amplitude. I_a is generally calculated from one of the horizontal components of the accelerogram. To eliminate a directional bias from one of the components being much larger than the other, we sum the respective I_a values of both components to produce I_b.

PREVIOUS USE OF ARIAS INTENSITY TO ESTIMATE LANDSLIDE EFFECTS

Using arguments regarding the critical acceleration thresholds for different general classes of landslides, WILSON & KEEFER (1985) calculated Arias intensity thresholds for landslides of different types. For rock falls and other disrupted failures, this was determined to be approximately $I_b = 0.25$ m/sec. Then, using seismograms from earthquakes in southern California, they calculated the distribution of this Arias intensity threshold in terms of magnitude and distance away from seismic source. Worldwide landslide data (KEEFER, 1984) plotted in terms of farthest distance from source versus magnitude was then compared with the magnitude-distance threshold plot of Arias intensity and found to be similarly distributed.

Having a theoretical relationship between Arias intensity, magnitude, and distance, we can then predict how far away from the seismic source that landslides are likely to occur for a given magnitude earthquake in terms of the probability of shaking exceeding the critical Arias intensity necessary for landslide initiation. Given a vertical strike-slip fault typical of the San Andreas system in California, the limits for the exceedance of critical Arias intensity levels are symmetrically located ellipses with respect to the source zone projected on the earth's surface. In reality, the locus of landslide limits from earthquakes are seldom symmetrical but are highly asymmetric with respect to the source because of variations in geologic materials and their respective susceptibility to failure and the variation in shaking caused by the seismic source itself and/or differences in shaking levels due to site response.

The first two earthquakes to give us real data to correlate Arias intensity values with landslide (rock fall) thresholds were the October 1, 1987 Whittier Narrows earthquake (M = 6.1) and the November 24, 1987 Superstition Hills earthquake (M = 6.6), both in southern California.

The magnitude 6.6 earthquake in the Superstition Hills produced in Arias intensity distribution that was, as we expected, less symmetrical with respect to the source area than theory predicted. However, when compared to the area limit of landslides, revealed an intensity range for the threshold of landslides that is consistent with the theory. The range of I_b for the triggering level of rock falls in this earthquake was $I_b \approx 0.2$-0.9 m/s.

The Whittier Narrows earthquake produced a much more complicated picture of the Arias intensity threshold for landslides. Here, there were two threshold ranges, one of $I_b \approx 0.42$-0.47 m/s and $I_b \approx 0.02$-0.07 m/s, the latter approximately an order of magnitude lower than that from the Superstition Hills earthquake.

Subsequent field investigation has established that the most obvious explanation for this dual threshold is that the threshold level of Arias intensity depends on site effects, particularly the fracture characteristics of the outcrops present. The lithology in areas of the higher threshold is made up of Miocene and Pliocene sandstone, shale, and conglomerate with relatively tight fractures. Although these deposits are relatively uncemented and present in steep slopes, the fractures having few large openings resulted in the higher Arias intensity threshold. In areas where the lower threshold occurred, the rocks ranged from Mesozoic sandstone and conglomerate to Precambrian (Archean) gneisses and schists with relatively loose open fractures and large quantities of loose rock as talus on slopes. Although these older rocks are harder, better cemented and more indurated, the large aperture of the fracture systems gives them a higher susceptibility to earthquake-induced rock fall and a lower threshold of Arias intensity.

As additional data becomes available from other earthquakes, we will be able to refine our data to more precise ranges of Arias intensity and have a more complete picture of its variations with respect to different rock types. We also anticipate defining Arias intensity levels for various general types of landslides.

SEISMICALLY INDUCED LANDSLIDES FROM THE LOMA PRIETA EARTHQUAKE

The M = 7.1, October 17, 1989, earthquake that struck the San Francisco Bay region triggered thousands of landslides over an area of 14,000 km². This region contains most of the San Francisco Bay area, the Santa Cruz and Monterey Bay area, nearby portions of the California Coast Ranges, and the coastline up to 130 km south of the epicenter. As well as causing over $10 million of damage to utilities, housing, and other structures, landslides blocked roads closing lifelines and hampering rescue efforts.

Landslides were most concentrated near the earthquake source in the heavily vegetated Santa Cruz Mountains, which had previously produced abundant landslides during years of heavy rainfall in winter storms. Landslides were numerous from this earthquake in spite of its occurrence at the end of a two-year drought period.

The most numerous types of landslides triggered were rock falls, rock slides, and soil slides that were typically smaller than 100 m^3 although some had volumes of between 1,000 and 10,000 m^3. Highway 17, between Los Gatos and Santa Cruz, a major avenue for commuters in the San Francisco Bay area, was closed for over a month by two rock falls.

Deeper seated block slides and rotational slumps, with failure surfaces possibly as deep as 70 m occurred in high concentrations along the south flank of Summit Ridge between Highway 17 and Highland Road. The top of Summit Ridge, which extends approximately East-West is occupied by Summit Road, a major access road for residents in the area. This and other roads as well as over 70 houses were heavily damaged by these landslides and other coseismic fissuring.

This area of deep-seated landslides and fissures lies immediately to the southwest of the surface trace of the San Andreas fault zone, and its center is approximately 10 km northwest of the earthquake epicenter. The network of fractures formed during the earthquake dissected the Summit Ridge area in a complicated way. The complex pattern produced by the landslides and other fractures presented a confusing picture to geologists who were unable to recognize primary surface faulting along the surface trace of the San Andreas fault itself. Instead of a clear zone of right lateral shear fractures, there was a zone up to 5 km wide in which fractures were left lateral, right lateral, extensile, normal, thrust, and combinations of these, mixed with what were, in some cases, clearly identifiable landslides.

A majority of the fractures with shear displacement showed left lateral offset. Fractures tended to be discontinuous. Fractures of several meters depth and up to 1.0 m of displacement seldom extended for more than 400 m. The majority of fractures showed extension to the dominant component of the total displacement. Fractures trending N40-80W tend to be parallel to bedding and/or regional faults. Therefore, fractures falling within this trend range are interpreted as being tectonic or related to regional structure rather than landslide related unless clearly associated with a landslide feature. Numerous other fractures form headwall scarps and lateral shear margins of deep-seated rotational slumps or block slides (landslide nomenclature is that of VARNES, 1978).

Fractures and fissures interpreted as being of tectonic origin include cracks located along linear ridge-top depressions and linear fractures that extend across well-defined landslide features. Some of these may have resulted from shear displacement along bedding planes of steeply dipping sandstone and shale formations (COTTON & *alii*, 1990). Other fissures may have formed as extensional partings along bedding or joints within the underlying sedimentary formations. The presence of uphill-facing scarps along such fractures suggests a tectonic origin, although this relationship is not in itself conclusive. Some uphill-facing scarps may have been formed by lateral deformation of the summit ridge with resultant subsidence of a ridge-top trough (sacküng).

Most recognizable among landslide-related cracks are extension fractures, commonly located along preexisting headwall scarps. These fractures have the largest displacements of all the landslide-related cracks. Displacements across the scarps of the largest deep-seated slumps exceed 1.0 m, are extensional and valley side down, similar to a normal fault.

Fractures that formed lateral margins of landslides generally exhibit shear displacement consistent with the relative displacement of the landslide mass. There are, however, numerous areas where the relative displacement of the landslide is not clear and where the relation of the sense of displacement across individual cracks to the overall mass movement is also unclear. The least common deformational features related to earthquake-triggered landslides are well-defined folds or fractures along the toes of landslides of which only a few isolated small features were observed.

Extensional fractures along the crests of narrow, steep-sided ridge tops occurred in several areas within the epicentral area. The intense ridge-crest fracturing occurred approximately 6 km West of the epicenter on a steep ridge where wood frame houses were located along the ridge crest. Shaking was so violent here that walls were torn from houses and, in some cases, complete collapse of the dwelling ensued. Rocks, concrete slabs, and logs in this area were also found displaced from their original positions with little or no evidence of transport by rolling. These objects appear to have been thrown from their original positions to their present ones by vertical accelerations of over 1.0 g.

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