Scale-invariance of soil moisture variability and its implications for the frequency-size distribution of landslides

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Power spectral analyses of soil moisture variability are carried out from scales of 100 m to 10 km on the microwave remotely-sensed data from the Washita experimental watershed during 1992. The power spectrum \( S(k) \) has an approximately power-law dependence on wave number \( k \) with exponent \(-1.8\). This behavior is consistent with the behavior of a stochastic differential equation for soil moisture at a point. This behavior has important consequences for the frequency-size distribution of landslides. We present the cumulative frequency-size distributions of landslides induced by precipitation in Japan and Bolivia as well as landslides triggered by the 1994 Northridge, California earthquake. Large landslides in these regions, despite being triggered by different mechanisms, have a cumulative frequency-size distribution with a power-law dependence on area with an exponent ranging from \(-1.5\) to \(-2\). We use a soil moisture field with the above statistics in conjunction with a slope stability analysis to model the frequency-size distribution of landslides. In our model landslides occur when a threshold shear stress dependent on cohesion, pore pressure, internal friction and slope angle is exceeded. This implies a threshold dependence on soil moisture and slope angle since cohesion, pore pressure, and internal friction are primarily dependent on soil moisture. The cumulative frequency-size distribution of domains of shear stress greater than a threshold value with soil moisture modeled as above and topography modeled as a Brownian walk is a power-law function of area with an exponent of \(-1.8\) for large landslide areas. This distribution is similar to that observed for landslides. The effect of strong ground motion from earthquakes lowers the shear stress necessary for failure but does not change the frequency-size distribution of failed areas. This is consistent with observations. This work suggests that remote sensing of soil moisture can be of great importance in monitoring landslide hazards and proposes a specific
I. INTRODUCTION

Landslides can be triggered by intense rainfall, rapid snowmelt, a change in water level, and ground motion from earthquakes and volcanic eruptions (Wieczoreck, 1996). The first three mechanisms act to increase soil moisture. The coincidences of increased groundwater levels and pore pressures with landslides have been investigated by many authors (Fukuoka, 1980). Studies suggest that a threshold dependence on soil moisture is appropriate for slope instability since landslides often occur when the product of rainfall duration and intensity exceeds a threshold value (Caine, 1980; Wieczoreck, 1987). In situ monitoring of pore pressure has shown that increases in pore pressure resulting from heavy precipitation is coincident with landslides (Johnson and Sitar, 1990). Several authors have shown that variations in the frequency of occurrence of landslides respond to climatic changes with higher rates of landslides associated with wetter climates (Grove, 1972; Innes, 1983; Pitts, 1983; Brooks and Richards, 1994). Further evidence for correlations between landslides and soil moisture is the observation that soil drainage is the most successful method of landslide prevention in the United States (Committee in Ground Failure Hazards, 1985).

Seismic triggering of landslides has also been studied by many authors (Keefer, 1984; Jibson, 1996). Correlations between landslides and earthquakes can be inferred in a number of ways. Many earthquakes have been known to trigger large numbers of landslides and landslide densities have a strong correlation with active seismic belts (Jibson and Keefer, 1988; Tibaldi et al., 1995). In some cases landslides have been associated with both high soil moisture and earthquakes (Seed, 1968).

Landslides are a serious natural hazard and also play an important role in the long-term shaping of landforms (Beck, 1968; Keefer, 1994; Densmore et al., 1997). In this paper we investigate the frequency-size distribution of landslides in areas where different triggering mechanisms dominate. We argue that fractal models are useful for modeling the
soil moisture and topographic variations that are associated with the landslide instability. A number of authors have examined the frequency-size distribution of landslides (Whitehouse and Griffiths, 1983; Ohmori and Hirano, 1988; Sugai, 1990; Sasaki, 1991; Hovius et al., 1997). These authors have consistently noted that the distribution is power-law or fractal. This is precisely analogous to the Gutenberg-Richter law for earthquakes which states that the cumulative frequency of events with a seismic moment greater than $M_o$ is a power-law function of $M_o$. In this paper we present cumulative frequency-size distributions for three new data sets. One of these data sets include landslides triggered exclusively by hydrologic mechanisms, one represents landslides triggered by a single large earthquake, and one includes landslides triggered by both hydrologic and seismic mechanisms. Surprisingly, despite different triggering mechanisms, the frequency-size distributions are very similar, with a power-law distribution for large areas and a flattening out of the distribution for small areas.

There is also observational evidence that soil moisture exhibits scale-invariant behavior. Rodriguez-Iturbe et al. (1995) have analyzed the frequency-size distribution of patches of soil moisture greater than a threshold value and have shown them to have a power-law cumulative frequency-size distribution. This is analogous to that for landslides. This suggests that soil moisture patches may be related to landslide failure areas.

In Section 2 of this paper we perform spectral analyses of soil moisture to show that soil moisture is a scale-invariant function. We show that the classic equation for soil moisture with evapotranspiration proportional to soil moisture and diffusive dispersion predicts the observed scale-invariance. In Section 3 we present frequency-size distributions of landslides and argue that a universal distribution exists, independent of triggering mechanism. Section 4 attempts to explain the observed frequency-size distribution of landslides with a slope stability analysis with soil moisture and topography modeled as scale-invariant functions.
II. POWER SPECTRAL ANALYSIS OF SOIL MOISTURE DATA

We have obtained microwave remotely-sensed soil moisture data collected from June 10 to June 18, 1992 at the Washita experimental watershed (Jackson, 1993). The watershed received heavy rainfall preceding the experiment but did not receive rainfall during the experiment. The data are gridded estimates of soil moisture in the top five centimeters of soil calculated using the algorithm of Jackson and Le Vine (1996). Each pixel represents an area of 200 m x 200 m and the total area considered is 45.6 km by 18.6 km. The soil moisture values do not correlate with relief in the watershed, which is very small. We have computed the one-dimensional power spectrum of each row of pixels for each image and averaged the power spectrum for each row at equal frequency values. The power spectrum $S$ is the square of the Fourier coefficients at each wave number of a Fourier series representation of the data. The power spectrum was estimated with the routine “spctrm” of Press et al. (1992). The spectra for June 10, 14, and 18 are presented from top to bottom in Figure 1. Although the soil moisture decreased significantly during the course of the experiment, the spectrum has a nearly constant form. The spectrum is given approximately by a power law $S(k) \propto k^{-1.8}$, also plotted in Figure 1. The power-law form with exponent $-1.8$ is chosen to compare with the model presented below. A power-law dependence of the power spectral density $S$ on wave number $k$, $S(k) \propto k^{-2}$, directly implies scale-invariant behavior. If $\beta = 2$ the behavior is a Brownian walk, with $\beta = 1.8$ it is a fractional Brownian walk. In both cases the behavior is that of a self-affine fractal (Turcotte, 1992). Although the spectrum appears to flatten out at high wave numbers, soil moisture variability at the Washita watershed with finer spatial resolution has been quantified by Rodriguez-Iturbe et al. (1995) using similar techniques. They find that the soil moisture variability continues to be scale invariant down to the 30 m scale for the more finely sampled data.

In the top image of Figure 2 we present a color image of the soil moisture at Washita on June 17, 1992. White spaces indicate areas where the watershed is interrupted by roads or lakes. Below it is a synthetic two-dimensional image constructed using the Fourier-filtering
method of Turcotte (1992). The mean and variance of the synthetic moisture field have been chosen to match that of the observed image and we have taken $\beta = 1.8$. The synthetic image reproduces the correlated structure of the real soil moisture image.

Entekhabi and Rodriguez-Iturbe (1994) have proposed a partial differential equation for the dynamics of the soil moisture field $s(\vec{x}, t)$:

$$\frac{\partial s(\vec{x}, t)}{\partial t} = D \nabla^2 s(\vec{x}, t) - \eta s(\vec{x}, t) + \xi(\vec{x}, t)$$

(1)

In this equation, the evapotranspiration rate $\eta$ is assumed to be constant in space and time. Soil moisture disperses in the soil according to a diffusion equation, and rainfall input $\xi(\vec{x}, t)$ is modeled as a random function in space and time. Variations in rainfall from place to place cause spatial variations in soil moisture that are damped by the effects of diffusion and evapotranspiration. Without spatial variations in rainfall input, there are no variations in soil moisture according to this model. As pointed out by Rodriguez-Iturbe et al. (1995), the variability at the small scale of the Washita watershed is not likely to be the result of spatial variations in rainfall.

An alternative approach to this problem which will generate spatial variations in soil moisture at these small scales assumes that the evapotranspiration rate is a random function in space and time. This models spatial and temporal variations in evapotranspiration resulting from variable atmospheric conditions and heterogeneity in soil, topography, and vegetation characteristics. The resulting equation is

$$\frac{\partial s(\vec{x}, t)}{\partial t} = D \nabla^2 s(\vec{x}, t) - \eta(\vec{x}, t)s(\vec{x}, t)$$

(2)

This equation is a variant of the Kardar-Parisi-Zhang equation (Kardar et al., 1986) which has received a great deal of attention in the physics literature. This equation is nonlinear and can only be solved numerically. Amar and Family (1989) have solved this equation and have found that the solutions are scale-invariant and fractal with a Hausdorff measure of approximately 0.4, independent of the relative strength of the diffusion and noise terms. This implies that the one-dimensional power spectrum has a power-law dependence on wave
The similarity between the field generated by equation (2) and the soil moisture observed in the Washita images suggests that equation (2) may be capturing much of the essential dynamics of soil moisture at these scales.

Rodriguez-Iturbe et al. (1995) quantified the scale-invariance of soil moisture variations by computing the cumulative frequency-size distribution of patches of soil with moisture levels higher than a prescribed value. They found that the cumulative frequency-size distribution had a power-law function on area with exponent of approximately $-0.8$. The number of soil patches $N$ depended on the area according to

$$N(> A) \propto A^{-0.8} \quad (3)$$

This is a fractal relation. We have determined the equivalent distribution for the synthetic soil moisture field illustrated in Figure 2. The result is plotted in Figure 3. The same distribution that was observed in real soil moisture fields is obtained. The same distribution has also been observed for cumulus cloud fields (Pelletier, 1997).

III. OBSERVED FREQUENCY-SIZE DISTRIBUTIONS OF LANDSLIDES

Fuyii (1969), Whitehouse and Griffiths (1983), Ohmori and Hirano (1988), Sugai (1990), Sasaki (1991), and Hovius et al. (1997) have all presented evidence that landslide frequency-size distributions are power-law functions of area. Some have presented cumulative distributions while others have presented noncumulative distributions. Care should be taken when comparing distributions to ensure that the same type of distribution is being used. In this section we present cumulative frequency-size distributions of landslide area from three areas. The results suggest that cumulative frequency-size distributions are remarkably similar despite different triggering mechanisms.
A. Landslides in Japan

A data set of 3,424 landslides with areas larger than $10^4$ m$^2$ in the Akaishi Ranges, central Japan, have been compiled by Ohmori and Sugai (1995). Landslide masses were selected for measurement from 1:20,000 scale aerial photographs and by field survey. On 1:25,000 scale topographic maps the length and area of each landslide was measured with a digitizer. In this region landslides occur as a result of both heavy rainfall and strong seismicity. The distribution of landslides is not uniform over the study area but correlates well with bedrock lithology. For this reason, the authors have divided their data set into 13 subsets each with a similar lithology. We have computed the cumulative frequency-size distribution of landslides in each of the lithologic units that contained at least 100 landslides. The data set is believed to be complete for landslides with areas greater than $10^{-2}$ km$^2$. In Figure 4 cumulative frequency-size distributions, the number of landslides with an area greater than $A$, are plotted as a function of $A$ for seven lithologic units. Areas for these landslides include the runout zone. The distributions plotted in Figure 4 are remarkably similar between the different lithologic units. For large landslides, the distribution is well characterized by a power law with an exponent of approximately $-2$, that is

$$N(> A) \propto A^{-2}$$

(4)

This power-law trend for large landslides was chosen based on visual correlation with the landslide data. Thus, the exponent $-2$ is only approximate. The observed frequency-area distribution flattens out for areas less than about $10^{-1}$ km$^2$, thus few landslides occur with small areas. Note that this change in the distribution occurs at an area that is an order of magnitude larger than the resolution of the catalog. Therefore, it does not appear to be an artifact of completeness of the data set.

It can be argued that a better estimate of the initial failure area is provided by the square of the width of the landslide since the length of the landslide includes the runout distance. The data set from Japan includes both estimates of total area as well as the
approximate length and width of each landslide. This allows us the opportunity to compare
the distribution of total landslide area to the distribution of the square of the width. In
Figure 5 the cumulative distributions of the total area and the width squared are plotted for
one of the lithologic units of the Japan data set. The upper line is the cumulative statistics
for total area and the lower line is the cumulative statistics for width squared. This figure
suggests that the total landslide area scales in approximately the same way as the initial
failure area. The two distributions were compared for the other lithologic units of the Japan
data set and results similar to those of Figure 5 were obtained.

B. Landslides in Bolivia

Two sets of landslide areas were mapped in adjacent watersheds in the Yungas region
of the Eastern Cordillera, Bolivia between Lake Titicaca and the city of Guanay. The main
river channels flowing through each of the watersheds are the Challana and an unnamed trib-
utary of the Challana. Because very few people inhabit this region, most of the landslides
(more than 95%) were not anthropogenically influenced. The watersheds are located in a rel-
atively aseismic region of the Andes; therefore, all of the landslides are thought to have been
hydrologically triggered during the wet season which extends from November through April.
Based on multitemporal aerial photographic coverage and limited dendrochronology field
studies, the landslide data sets represent an interval of time of about 20-35 years. Scanned
aerial photographs at approximately 1:50,000 scale were used for mapping. Each photo-
graphic image was georegistered to a Thematic Mapper image using a Universal Transverse
Mercator projection. Landslide mapping was only undertaken in suitable portions of the
watershed below treeline. This excluded cloud covered, shadowed, and agricultural areas.
The full extent of each landslide was mapped including the runout zone. Since individual
trees can be discerned in the mapped areas, the resolution of the photographs is about 3 to
10 meters.

The cumulative frequency-size distributions of landslides in these data sets are presented
in Figure 6. Distributions similar to those obtained with the data from Japan were observed. Large landslides in the two areas match power-law distributions from equation (4) with exponents $-1.6$ and $-2$, determined by visual correlation. A flattening of the distribution at small areas was also observed. As with the Japan data set, this rolloff is at a scale much larger than the resolution of the data set.

C. Landslides triggered by the 1994 Northridge, California earthquake

The January 17, 1994 Northridge, California earthquake triggered more than 11,000 landslides over an area of 10,000 km$^2$. Harp and Jibson (1995) mapped landslides triggered by the earthquake. They used 1:60,000-scale aerial photography taken the morning of the earthquake by the U.S. Air Force. The subsequently digitized photos were supplemented by field work. It was estimated that the inventory is nearly complete down to landslides of about 5 m on a side below which a significant number may have been missed. The distribution is plotted in Figure 7. Large landslides have a power-law dependence on area with an exponent of approximately $-1.6$. Again, there is a rolloff for small areas. As with the Bolivian and Japanese data sets, this rolloff is at a scale much larger than the resolution scale of the data set and is not an artifact of an incomplete catalog.

IV. MODELING OF LANDSLIDE FAILURE WITH REALISTIC TOPOGRAPHY AND SOIL MOISTURE

Landslides occur when the shear stress exceeds a threshold value given approximately by the Coulomb failure criterion (Terzaghi, 1962)

$$\tau_f = \tau_0 + (\sigma - u) \tan \phi$$

(5)

where $\tau_0$ is the cohesive strength of the soil, $\sigma$ is the normal stress on the slip plane, $u$ is the pore pressure, and $\phi$ is the angle of internal friction. Landslides are initiated in places where $\tau_f$ is greater than a threshold value. The movement of the soil at the point of
instability increases the shear stress in adjacent points on the hillslope causing failure of a connected domain with shear stress larger than the threshold value. To model the landslide instability and, in particular, the frequency-size distribution of landslides, it is therefore necessary to model the spatial variations of $\tau_0$, $\sigma$, $u$, and $\phi$. The variables $\tau_0$ and $u$ are primarily dependent on soil moisture for a homogeneous lithology and a slip plane of constant depth. The dependence of each of these variables on soil moisture has been approximated using power-law functions (Johnson, 1984). The shear stress and normal stress are linearly proportional to soil moisture through the weight of water in the soil. The shear stress and normal stress are also trigonometric functions of the local slope. Based on the results given in Section 2 the soil moisture will be modeled as a two-dimensional fractional Brownian walk with $\beta = 1.8$. Topography will be modeled as a Brownian walk. Power spectral analyses of one-dimensional transects of topography have shown topography to be a scale-invariant function analogous to soil moisture. The power spectrum has a power-law dependence with an exponent $\beta = 2$ in a variety of tectonic and erosional environments (Mandelbrot, 1975; Sayles and Thomas, 1978; Turcotte, 1987; Huang and Turcotte, 1989). Thus it is a Brownian walk. This scale-invariant behavior is applicable only over a certain range of scales. At scales smaller than the support area of a drainage network, for instance, the surface is not dissected by channels and the hillslope is smooth. To model the small-scale smoothness of topography as well as the scale-invariant behavior at larger scales, we have constructed a two-dimensional surface with one-dimensional power spectrum $S(k) \propto k^{-2}$ with the Fourier-filtering method of Turcotte (1992). At small scales, the synthetic topography has been planarly interpolated. A shaded-relief example of the model topography is illustrated in Figure 8 along with its contour plot. The plot has 128 x 128 grid points with interpolation below a scale of 8 pixels. The result of the interpolation is clearly identified as piecewise linear segments in the transects along the boundaries of the plot. The slope corresponding to this model of topography is illustrated in Figure 9. Below a scale of 8 pixels, the slopes are constant. Above this scale, the slopes are a two-dimensional Gaussian white noise. This follows from the fact that our model for topography at large scales, a Brownian walk with $S(k) \propto k^{-2}$,
can be defined as the summation of a Gaussian white noise time series. White noise means that adjacent values are totally uncorrelated. The contour map of the slope function shows that there are no contours smaller than 8 x 8 pixels.

The shear stress necessary for failure is a complex function of soil moisture and slope. However, to show how landslide areas may be associated with areas of simultaneously high levels of soil moisture and steep slopes, we will assume a threshold shear stress criterion proportional to the product of the soil moisture and the slope. In addition, we will assume that slope and soil moisture are uncorrelated. A grid of synthetic soil moisture and topography of 512 x 512 grid points was constructed according to the models described above. The domains where the product of the soil moisture and the topography were above a threshold value are shown in Figure 10. The threshold value was chosen such that only a small fraction of the region was above the threshold. Figure 11 shows the cumulative frequency-size distribution of the regions above threshold, our model landslides. It can be seen that at large areas a power-law distribution with an exponent of −1.6, similar to that observed for landslides, is observed. Distributions obtained with different realizations of the synthetic random fields had exponents of 1.6 ± 0.1 fit to the landslides above A = 10. For values of the threshold that resulted in only a small fraction of the lattice being above threshold, the form of the distribution was independent of the value of the threshold. The exponent of this distribution is more negative than that of the soil moisture patches of Figure 3. This results from the less correlated slope field “breaking up” some of the large soil moisture patches so that there are fewer large landslides relative to small ones than there are large soil moisture patches relative to small ones. The effect of the smooth topography at small scales acts to flatten the frequency-size distribution below the power-law trend. This means that the slope function is completely correlated at these scales. The slope tends to fail as a unit more often than it would if the slope was not constant on these scales. This results in fewer small avalanches and a flattening out of the distribution for small areas. The flattening out of the observed landslide distributions at small areas can also be associated with a minimum thickness necessary for failure at depth.
The effect of strong ground motion from earthquakes is to lower the shear stress necessary for failure (Newmark, 1965). This does not alter the frequency-size distribution of landslides according to our model since the form of the distribution is independent of the value of the threshold. This is consistent with the observation of Section 3 that the cumulative frequency-size distribution is independent of triggering mechanism.

V. DISCUSSION

There has been a great deal of interest in the physics community in systems exhibiting “avalanches.” Much of this interest stems from the theory of self-organized criticality recently introduced by Bak et al. (1988). These authors proposed the idea that many systems are driven to a state in which energy is dissipated in events which have a power-law distribution of sizes. The type example of such a system is a growing sandpile. Sand grains are dropped on a pile continuously and are lost from the edge of the pile in discrete sand slides. A few sand slides are large and many are small. Experimental studies of real sand piles suggest that the frequency-size distribution of these slides may in fact be power-law (Held et al., 1990; Bretz et al., 1992; Rosendahl et al., 1994; Densmore et al., 1997). Rothman et al. (1993) has presented evidence for power-law statistics in the cumulative frequency-size distribution of thicknesses of turbidites. It should also be noted that they observed a flattening of the distribution at small thicknesses similar to that which we have observed for small areas in the frequency-size distribution of landslides. The results of Rothman et al. (1993) have been interpreted in terms of the theory of self-organized criticality. Self-organized criticality has also been proposed as model for earthquakes (Bak and Tang, 1989) and landslides (Noever, 1993). The model we have presented appears to be an alternative hypothesis for the dynamics of landslides. The behavior of our model is a result of soil moisture dynamics that are independent of topography. In the sandpile model self-organization of the sandpile into a critical state results from the continuous input of energy into the system and the action of large and small sand slides. In that model there is no additional field that affects
the system behavior. In contrast, in our model soil moisture acts as an additional field that affects the system dynamics.

Since soil moisture is known to be an important factor in the landslide instability and large landslides obey power-law statistics, it is likely that patches of soil moisture above a threshold value, which also obey power-law statistics, can be associated with landslides through the model we have discussed. This observation suggests that continuous remote-sensing of soil moisture, together with a digital elevation model, are necessary for successful landslide hazard assessment. A continuous monitoring program of landslide-prone areas could also provide valuable data for the better understanding of landslide failure. For example, in the model we presented, landslide failure was chosen to be the product of slope and soil moisture. As noted, the correct failure criterion is likely to be a complex function of the slope and the soil moisture. This function could be determined empirically using historical records of landslides if high resolution soil moisture data and topography are available for the time immediately preceding each landslide.

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FIGURE CAPTIONS

Figure 1: Average one-dimensional power spectrum of soil moisture on (top) June 10, (middle) June 14, and (bottom) June 18, 1992 in the Washita experimental watershed. The spectra are offset so that they can be plotted on the same graph.

Figure 2: (a) Color map of microwave remotely-sensed estimates of soil moisture on June 17, 1992 at the Washita experimental watershed. (b) Color map of a synthetic soil moisture field with the power spectrum $S(k) \propto k^{-1.8}$ with the same mean and variance as the remotely-sensed data.

Figure 3: Cumulative frequency-size distribution of patches of soil moisture larger than a threshold value for the synthetic soil moisture field with power spectrum $S(k) \propto k^{-1.8}$.

Figure 4: Cumulative frequency-size distribution of landslides in 6 lithologic zones in Japan. The distributions are well characterized by $N(> A) \propto A^{-2}$ above approximately 0.1 km$^2$.

Figure 5: The upper curve is the cumulative frequency-size distribution for one of the lithologic zones in Japan as illustrated in Figure 4. The lower curve is the data for the same set of landslides with the area defined to be the width of the landslide squared.

Figure 6: Cumulative frequency-size distribution of landslides in two areas in Bolivia.

Figure 7: Cumulative frequency-size distribution of landslides triggered by the 1994 Northridge, California earthquake.

Figure 8: Synthetic model of topography used in the model of Section 4. The topography is a scale-invariant function with $S(k) \propto k^{-2}$ above a scale of eight lattice sites and is planarly interpolated below that scale.

Figure 9: Slope model corresponding to the topographic model of Figure 8. The slopes are white noise above a scale of 8 lattice points and constant below that scale.

Figure 10: Contour map of the product of the synthetic soil moisture field with the synthetic slope function. Areas inside the contour loops represent model landslides.

Figure 11: Cumulative frequency-size distribution of model landslides. The distribution
compares favorably to the distributions of real landslides.
