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Topographic controls of landslides in Rio de Janeiro: field evidence and modeling

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Abstract

Landslides are common features in the *Serra do Mar*, located along the southeastern Brazilian coast, most of them associated with intense summer storms, specially on the soil-mantled steep hillslopes around Rio de Janeiro city, where the *favelas* (slums) proliferated during the last few decades. On February 1996, hundreds of landslides took place in city of Rio de Janeiro triggered by intense rainstorms. Since then, many studies have been carried out in two experimental river basins in order to investigate the role played by the topographic attributes in controlling the spatial distribution of landslides inside them. Landslide scars and vegetation cover were mapped using aerial photographs and field observations. A detailed digital terrain model (4 m^2 resolution) of the basins was generated from which the main topographic attributes were analyzed, producing maps for slope, hillslope form, contributing area and hillslope orientation. By comparing these maps with the spatial distribution of the landslide scars for the 1996 event, a landslide potential index (LPI) for the many classes of the different topographic attributes was defined. At the same time, field experiments with the Guelph permeameter were carried out and a variety of scenarios were simulated with the SHALSTAB model, a process-based mathematical model for the topographic control on shallow landslides. The results suggest that most of the landslides triggered in the studied basins were strongly influenced by topography, while vegetation cover did affect landslide distribution. Between the topographic attributes, hillslope form and contributing area played a major role in controlling the spatial distribution of landslides. Therefore, any procedure to be used in this environment towards the definition of landslide hazards need to incorporate these topographic attributes.

Keywords: Landslide hazard; Hillslope hydrology; Mathematical modeling; Tropical soils

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1. Introduction

Landslides are common processes in the *Serra do Mar*, a long coast range that follows almost all the way from south to southeastern Brazil, well-known by its steep scarps that lead to the generation of fantastic landforms like the sugar-loaves. In the major cities, particularly the city of Rio de Janeiro, these processes triggered by intense summer rainstorms, have become more frequent since the 1960s (Meis and Silva, 1968a,b; Barata, 1969; Costa Nunes, 1969; Jones, 1973), when the occupation quickly spread towards the steep hillslopes inside and around the city. Nowadays, many of the soil-mantled steep hillslopes are densely occupied (Fig. 1), especially by the slums (*favelas*), affecting slope stability by the extensive usage of cuts, landfills, deforestation, changes in drainage conditions, accumulation of trash deposits on hillslopes, among others, adding new relationships to the natural conditioning factors related to geology and geomorphology (e.g., Brunsden and Prior, 1984; Sidle et al., 1985; Crozier, 1986). However, along the steep slopes of the Serra do Mar escarpments in Southeastern Brazil, it is evident that landslides play a major role in controlling landscape evolution in the long term (e.g., Bigarella et al., 1965; Meis and Silva, 1968a; Modenesi, 1988; Lopes, 1997; Furian et al., 1999; Coelho Netto, 1999; Cruz, 2000).

A recent landslide inventory of Rio de Janeiro city (Amaral and Palmeiro, 1997b), although covering just the period from 1962 to 1992, attests an increase in slope instability during the years of 1967, 1986 and 1988, directly related to intense summer rainstorms registered in those years. A future update in this inventory will certainly include 1996 in this list. From 1986 to 1996, landslide disasters killed 123 people and destroyed 414 houses in the city (Amaral, 1997). It is evident that the development of procedures able to carry out effective landslide prediction in such areas, although urged, is not easily achieved.

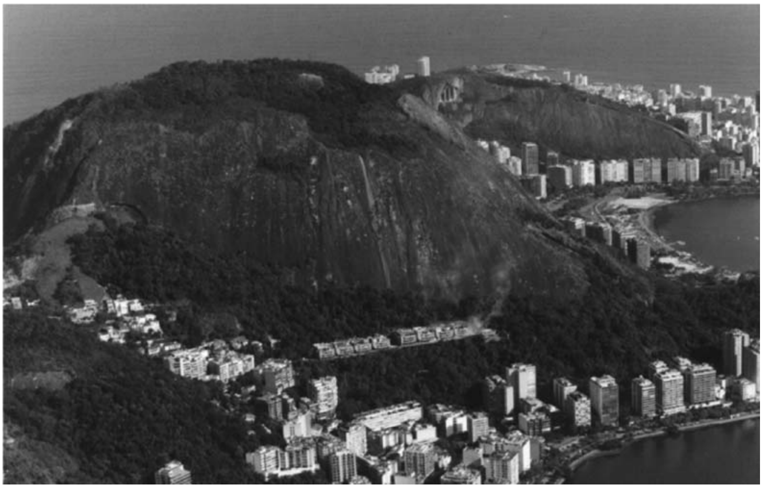

Fig. 1. Spreading of occupation towards the hillslopes around Lagoa Rodrigo de Freitas, Rio de Janeiro city.

Intense summer rainstorms are the main landslide-triggering factor. Unfortunately, rainfall amounts in the order of 200–300 mm in just 1–2 days are not rare in the city (Brandão, 1997). For example, in January 1966, a summer storm dumped about 480 mm rainfall over downtown Rio de Janeiro (about sea level) in just 3 days, while at stations placed at higher elevations (about 500 m), rainfall values got closer to 700 mm during this short period (Jones, 1973). More recently, in February 1988, about 384 mm rainfall were accumulated in the city during 4 days, with half of it registered during just one night (Brandão, 1997).

Geological features associated with the Pre-Cambrian bedrocks (granites, biotite and plagioclase gneisses, migmatites, basic and alkaline dikes, etc.), including metamorphic foliation, unloading and tectonic fractures, also play a major role in landsliding within the city limits (e.g., Barata, 1969; Jones, 1973; Barros et al., 1988; Amaral et al., 1992; Barroso, 1992; Amaral and Palmeiro, 1997a). Unloading fractures, for example, control the downward migration of the weathering processes (differential weathering), resulting in zones of similar weathering inside the thick weathering mantles, attested in many places by abrupt soil-bedrock boundaries. Frequently, these fractures can be clearly observed within the city, representing important mechanical and hydrological discontinuities.

Slope deposits are also directly associated with landslides in Rio de Janeiro. On the foothills of the escarpments, as well as in most of the topographic hollows, highly heterogeneous talus deposits can be observed. In some hillslopes, usually associated with saturated soils, these deposits can move very fast, attaining rates of 12.7 mm/day, as observed by Lacerda (1997) in the Tijuca massif. Colluvial deposits in the city, although thinner than the ones observed in other places in southeastern Brazil, may locally be important in defining critical places, particularly when these deposits are layered (Fernandes et al., 1994).

The colluvial mantle along the hillslopes of the city has a high clay content, usually above 40%, overlaying a thick sandy-silt saprolite. As already suggested by Jones (1973), and later confirmed by many others, the hydraulic conductivity in this environment tends to be greater in the saprolite than in the colluvial mantle (e.g., Wolle and Hachich, 1989; Vieira and Fernandes, 2003). Such behaviour has important implications for defining the rupture mechanisms in these type of hillslopes, particularly those concerned either with the development of critical, positive pore-water pressures inside saturated soils or with the loss of soil suction and decrease in apparent cohesion inside unsaturated soils (e.g., Vargas et al., 1986; Wolle and Hachich, 1989; Campos et al., 1994).

By comparing the location of landslide scars, associated with one single event or with a sequence of events, with the spatial distribution of the topographic attributes observed in the field, one can estimate the effect of each parameter on landsliding, as well as the triggering mechanisms (e.g., Gao, 1993; Larsen and Torres-Sanchez, 1998; Montgomery et al., 2000; Zhou et al., 2002). Besides, they improve our prediction abilities. These analyses have become more widely accessible due to increasing availability of high-quality digital elevation models.

In general terms, previous studies concerned with the role played by geomorphology in landsliding in southeastern Brazil have focused their attention mainly to the steepness parameter. For example, Cruz (1974) estimated threshold slope angles of about 22° (40%)

for the Serra do Mar escarpments in Caraguatatuba, south of Rio de Janeiro in São Paulo State, by overlaying the spatial distribution of landslide scars with the map for slope angles. Similar values (from 20° to 29°) were later obtained by field experiments with a portable shear device in the same area (Cruz and Colangelo, 2000). Threshold angles have also been estimated in other places around the world. For example, Gao (1993) observed that the landslide potential increased rapidly for hillslopes above 31° in Virginia (USA), while Zhou et al. (2002) observed that most of the landslides in Hong Kong, following the 1993 event, took place on slope angles above 25°–30°.

Although slope is a major topographic attribute, affecting both the hydrological conditions and stability analysis, its importance seems to have been overestimated in landslide hazard mapping procedures, particularly in Rio de Janeiro. Consequently, gentle hillslopes initially considered as having low landslide susceptibility were affected by landslides, especially debris-flows, during the 1996 rainstorms. This attests that other topographic parameters must be taken into consideration.

In Southeastern Brazil, few studies have tried to consider the contribution of other morphological attributes on landsliding. Hillslope form, for example, although earlier suggested as an important parameter (Meis and Silva, 1968a,b), was not incorporated into stability analysis and landslide hazard mapping procedures. More recently, the role played by concave forms (hollows) has been intensively investigated, including their effects on surface and subsurface hillslope hydrology (e.g., Coelho Netto, 1985; Fernandes et al., 1994), on gullying (e.g., Coelho Netto et al., 1988; Coelho Netto and Fernandes, 1990; Coelho Netto, 1999), as well as on landsliding (e.g.; Lacerda, 1997; Guimarães et al., 1999; Guimarães, 2000; Fernandes et al., 2001).

Another important topographic attribute is contributing area (per unit contour) since it defines the location of the convergent segments in a landscape. These portions are directly associated with the concentration of surface and subsurface flows, contributing to the development of soil saturation (e.g., Beven and Kirkby, 1979; O'Loughlin, 1986) and many studies have attested the importance of contributing area in defining critical areas for landsliding (e.g., Dietrich et al., 1993, 1995; Montgomery and Dietrich, 1994; Wieczorek et al., 1997; Montgomery et al., 2000).

The effective use of mathematical models for landslide prediction implies the understanding of the complex interactions between triggering mechanisms and conditioning factors. This knowledge requires, on the other hand, detailed field data concerning soil properties (cohesion, friction angle, specific weight, hydraulic conductivity, etc.) and hillslope hydrology, particularly the spatial and temporal variations in pore-water pressures (e.g., Anderson and Burt, 1978; Coelho Netto, 1985; Wilson and Dietrich, 1987; Harp et al., 1990; Fernandes et al., 1994; Montgomery et al., 1997; Gerscovich et al., 1997).

In February 1996, hundreds of landslides were triggered in the city of Rio de Janeiro after a series of intense rainstorms, especially inside the Tijuca Massif (Fig. 2). In some parts of the city, during just 2 days, more than 350 mm of rain was recorded (GEORIO, 1996; Brandão, 1997; Coelho Netto, 1999). Most of the mass movements were shallow translation slides that converged towards the main valleys leading to catastrophic debris flows, which were able to move huge blocks and generated a great amount of destruction, including the loss of 222 houses and the death of 44 people (Amaral, 1997).

Fig. 2. Shallow landslides triggered by intense rainstorms on February 1996 on the southwestern flanks of the Tijuca Massif.

The most catastrophic debris-flows triggered in this event, which alone destroyed about 150 houses (Amaral, 1997), were located at Quitite and Papagaio river basins, draining the southwestern flanks of the Tijuca Massif towards the Jacarepaguá lowlands (Fig. 3). In these two basins (5 km^2), around 100 landslide scars were mapped in the 1996 event and since then a series of studies have been carried out inside their limits (e.g., Vieira et al., 1997; Guimarães et al., 1999, 2003; Gomes, 2002; Vieira and Fernandes, 2003).

In this study, we combined field evidences with a process-based mathematical model (Shallow Stability—SHALSTAB) in order to investigate the spatial relationships between landslides and their conditioning factors in Rio de Janeiro city. The study was applied to the Quitite and Papagaio basins and emphasis was given to the role played by topography, vegetation cover, and land-use in controlling the spatial distribution of landslide scars during this event. The ability of the model to predict landsliding in these basins was evaluated by comparing the location of the predicted unstable areas with the location of the actual scars for the 1996 event.

2. Study area

The two studied basins, Quitite and Papagaio, draining side by side the west flank of the Tijuca massif, have areas of 2.13 and 2.22 km^2, respectively, and about 100 landslide scars were mapped in the 1996 event (Fig. 4). Since then, a series of studies has been carried out in these basins (e.g., GEORIO, 1996; Vieira et al., 1997; Macias et al., 1997; Guimarães et al., 1999, 2003; Gomes, 2001; Vieira and Fernandes, 2003).

Elevations decrease from 975 to 20 m in about 4 km, where the rivers start to flow within densely occupied lowlands of Jacarepaguá. In the upper portions of the basins, steep forested slopes prevail and thin soils are usually less than 2.0 m thick, while in the

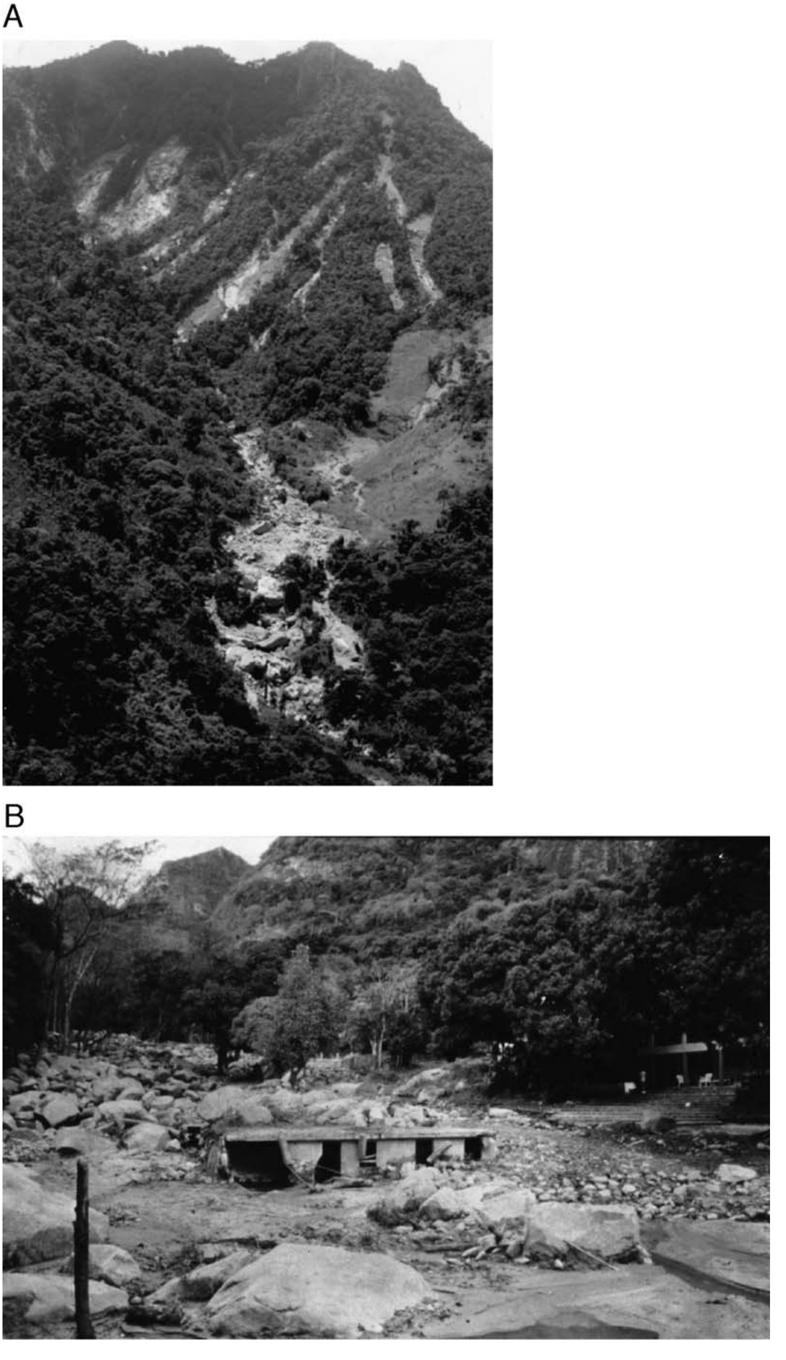

Fig. 3. Upper A and lower B portions of the Quitite debris-flow (Tijuca Massif) which destroyed many houses and other buildings.

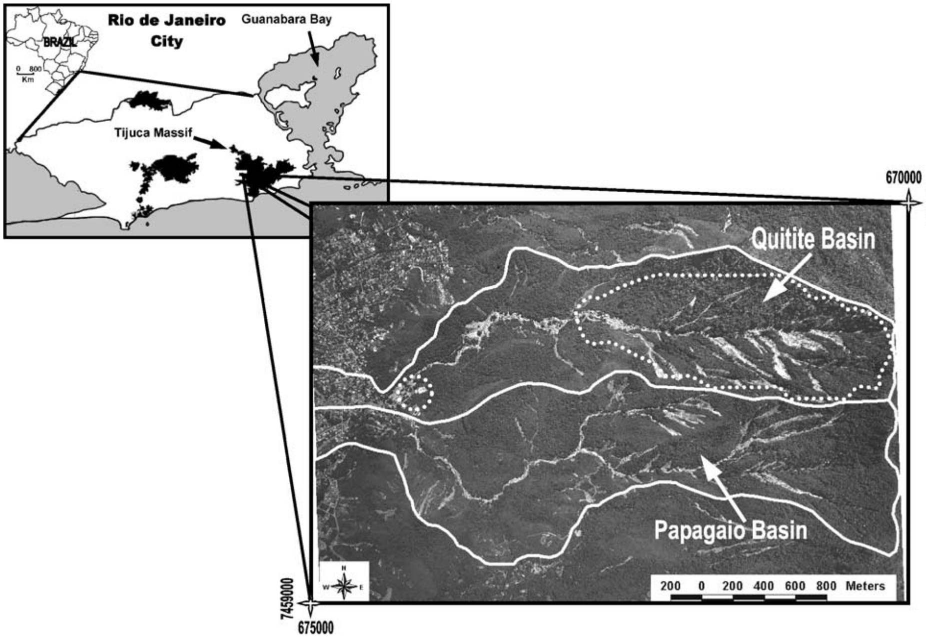

Fig. 4. Location of two studied basins (Quitite and Papagaio) in the southwestern flank of the Tijuca Massif. This aerial photograph shows the landslide scars from the 1996 event (white) in the upper parts of the basins and the debris-flows in the lower ends. About 150 houses were partially destroyed in the area where the two debris-flows converge.

middle portions part of the forest has already been substituted by grasslands. At that point in the basins, slopes become less steep and soil thickness frequently attains about 4.0–5.0 m. In the lower portions of the basins, densely occupied nowadays, the channels have been modified and houses constructed over previous river beds and the valleys are filled with old debris-flow deposits. Drillings on these deposits showed that they can be thicker than 12.0 m (GEORIO, 1996), attesting the recurrence of debris-flow processes in this region.

In general terms, bedrock is composed of a complex combination of Pre-Cambrian high-grade metamorphic rocks with granite intrusions (GEORIO, 1996; Coelho, 1997). The most frequent lithologic unit is a highly foliated banded gneiss, (Archer Gneiss). Although the foliation planes are variable, the predominant directions are between 000°/30° and 180°/30° (direction of the dip/dip angle). A medium texture, highly homogeneous granite intrudes the previous unit forming tabular levels with variable thickness, particularly in the upper portions of the basins. Different types of fracture sets are observed in the area, including unloading fractures and sub-vertical tectonic fractures with prevailing directions between 300° and 340°, and with a second set between 050° and 070°. Many dikes, mostly with a basic composition, are parallel to these directions and have affected landscape evolution by defining preferential sites for drainage incision.

3. Methods

3.1. Field mappings and landslide scars

Landslide scars were mapped in 1:20.000 aerial photographs, taken just 2 months after the 1996 event. Most of them were later visited in the field in order to define landslide types, soil characteristics, thickness of the failed material, soil depth and geology. Field mapping was also carried out to generate drainage and vegetation maps (Vieira et al., 1997, 1998). Detailed geologic mapping has also been conducted in some parts of the two basins (GEORIO, 1996; Coelho, 1997), but access to the steep and dense forested slopes in the upper portions is very difficult.

3.2. Topographic attributes and landslide potential indices

A detailed digital elevation model, with a 2.0×2.0 m grid, was generated for the two basins from the restitution of the aerial photographs described before. The spatial distribution of the investigated topographic attributes (slope, hillslope form, contributing area and hillslope orientation) was mapped using a geographic information system (GIS). These maps, together with the vegetation map, were overlaid with the landslide scar map, allowing the definition of the classes prevailing in each cell inside the landslide scars, for the four topographic variables, as well as for the vegetation cover.

Frequency maps for the classes considered for the four topographic attributes were then produced. A landslide potential index (LPI) was obtained for each terrain variable by calculating the ratio between the number of cells disturbed by landslides and the total number of cells for that specific class. This index identifies the relative influence of each class on landsliding, for vegetation and for each one of the four topographic attributes investigated. More details concerning these procedures can be obtained in other studies (Vieira et al., 1998; Guimarães et al., 1999; Guimarães, 2000).

3.3. Modeling landslide susceptibility

We applied the model SHALSTAB (Dietrich and Montgomery, 1998) to the Quitite and Papagaio basins, simulating a variety of scenarios (Guimarães et al., 1999, 2003; Guimarães, 2000). Because this model simulates the topographic control on shallow landsliding, it was considered suitable to be used in the study area. During the numerical simulations, the topographic variables were obtained from the detailed digital elevation model of the basins while the soil properties were estimated from field investigations and from previous studies carried out in areas nearby. Although the model allows the incorporation of the spatial variability of soil properties, in this study, they were considered constant within the two basins. Other studies are being conducted in order to analyze these influences on landsliding.

The SHALSTAB model is a deterministic mathematical model, which defines the relative susceptibility to shallow landsliding in a defined landscape. It has been developed since the early 1990s (Dietrich et al., 1992, 1993, 1995; Montgomery and Dietrich, 1994) and has been applied to many sites in the west coast of the United States (e.g., Montgomery et al., 1998, 2000; Dietrich and Sitar, 1997). This model assumes that although site specific properties control the size and the moment when shallow landslides are triggered, the main controlling factor defining their location is topography (Montgomery and Dietrich, 1994). Based on this assumption, SHALSTAB combines a steady state runoff model that estimates the topographically induced spatial variation in pore pressures with an infinite slope model for shallow landsliding (Dietrich and Sitar, 1997). The calculations incorporate, for each cell in the basin, variables associated with topography (slope and contributing area), usually obtained from the digital elevation model, climate (precipitation), as well as soil properties (thickness, hydraulic conductivity, density, cohesion and friction angle).

A landslide inventory of Rio de Janeiro (Amaral and Palmeiro, 1997b), considering all the landslides that occurred in the city from 1962 to 1992, has shown that shallow translation landslides are the most important in the city, corresponding to about 38% of all landslides. Other studies attest that they also predominate in other parts of the Serra do Mar (e.g., Cruz, 1974; De Ploey and Cruz, 1979; Wolle and Carvalho, 1989; Lacerda, 1997). Therefore, due to their regional importance, efficient procedures, able to predict the spatial distribution of shallow landslides in this tropical environment, are urged. Because we hypothesize that topography, particularly contributing area (per contour width), plays a major role in controlling landsliding in the Serra do Mar, the SHALSTAB model seems to represent a useful tool for regional landslide prediction in such an environment.

Eq. (1) shows the mathematical expression SHALTAB computes for each cell in the grid (4 m^2, in this study). Although the equation can be solved for the critical rainfall (Qc) required to trigger landslides in the study area, since we did not have many reliable data concerning the spatial variability of soil transmissivity (T), we solved it for the ratio Q_c/T, as suggested by Dietrich and Montgomery (1998). For the different scenarios analyzed, all the cells in the grid were classified according to the computed values for the ratio Q_c/T. The greater its absolute value, the more unstable the hillslope is. More details on this analysis can be obtained elsewhere (e.g., Montgomery and Dietrich, 1994; Dietrich and Montgomery, 1998).

$$\frac{Q_c}{T} = \frac{\sin\theta}{(a/b)}\left[\frac{C'}{\rho_w g z \cos^2\theta \tan\phi} + \frac{\rho_s}{\rho_w}\left(1 - \frac{\tan\theta}{\tan\phi}\right)\right] \tag{1}$$

where: Q_c is the critical rainfall required to fail, T is the soil transmissivity (the product between soil thickness and saturated hydraulic conductivity), a/b is the contributing area per contour width, θ is the local slope, ρ_w is the density of the water, g is the acceleration of gravity, z is soil thickness, ρ_s is the soil bulk density, ϕ is the soil friction angle and $C'$is the soil cohesion.

3.4. Hydraulic conductivity

Soil hydraulic conductivity plays an important role in landsliding, especially in areas with thick weathered profiles, like most of the tropical hillslopes. Differential weathering usually generates hydraulic discontinuities inside the weathered mantle, which may act as impeding layers during infiltration of intense summer rainstorms. These layers may locally contribute to soil saturation and to the development of critical pore-water pressures. Besides, as suggested by other studies carried out in the Serra do Mar (De Ploey and Cruz, 1979), steep forested hillslopes may present, as observed in Caraguatatuba (São Paulo), soils with high hydraulic conductivity values, allowing rapid infiltration and interflow along soil cracks and roots, which contribute to landsliding in this environment.

Other studies in southeastern Brazil have shown that, in contrast to the typical weathered profile of temperate region, which shows a decrease in saturated hydraulic conductivity with depth, thick tropical soils may show instead an increase with depth (Wolle and Hachich, 1989; Wolle and Carvalho, 1994). The pattern of changes in hydraulic conductivity with depth will directly affect slope stability because it will define the establishment of either lateral parallel flow or vertical downward flow.

A variety of methods have been proposed to estimate soil hydraulic conductivity in the field (e.g., Freeze and Cherry, 1979; Stephens, 1996; Tindall and Kundel, 1999). In this study, we used the Guelph permeameter (Reynolds and Elrick, 1985) to estimate vertical and lateral variations in hydraulic conductivity (both for the soil and for the weathered mantle) for different bedrock types that outcrop in the Papagaio basin. To investigate the hypothesis that hydraulic discontinuities inside the weathered mantle may have locally contributed to landslide generation in the basin, field measurements of hydraulic conductivity were carried out close to the border and inside landslide scars from the 1996 event. About 90 measurements of the saturated hydraulic conductivity were obtained, comprising the entire profile from the surface to bedrock, or to a maximum depth of about 6.0 m (Vieira and Fernandes, 2003). Because of the portability of the Guelph permeameter, some experiments were conducted close to landslide scars located on steep slopes at elevations above 500 m.

4. Results and discussion

4.1. Topographic attributes, vegetation cover and landsliding

Fig. 5 shows the frequency distribution and the landslide potential index (LPI) for the four topographic attributes studied (slope, hillslope form, contributing area and hillslope orientation) as well as for the vegetation cover. The results attest that slope angles between 18.6° and 37.0° are the most frequent to fail in the two basins (Fig. 5a). They also show that the LPI increases when slope angles increase until a limiting slope class of 37.1°–55.5° is achieved. For angles above this limit, LPI decreases attesting that these very steep hillslopes were less disturbed by landslides than others more gentle. In our basins, the steepest sites are associated with shallow soils, which may have already previously failed, and bedrock escarpments. These results, suggesting the existence of threshold slope angles for landsliding, are similar to the ones presented by Gao (1993), Larsen and Torres-Sanchez (1998) and others.

Hillslope form (Fig. 5b), as already expected, played a major role in controlling landslide distribution in the two basins. Although convex hillslope forms are the most

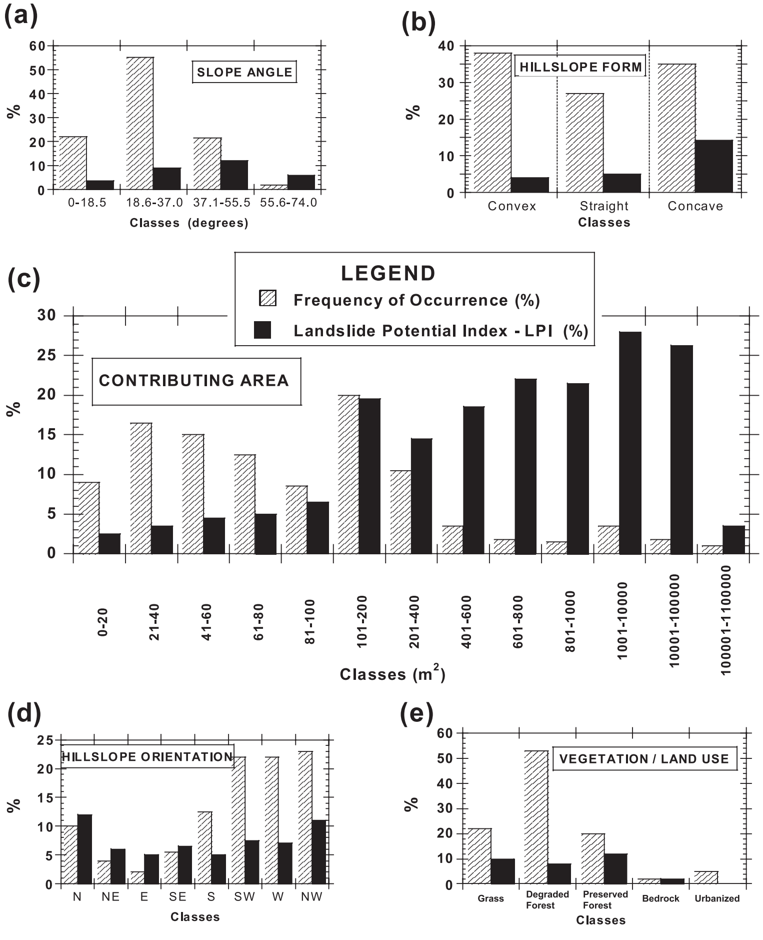

Fig. 5. Frequency distribution and the landslide potential index (LPI) for the topographic attributes slope angle (a), hillslope form (b), contributing area (c), hillslope orientation (d), as well as for vegetation/land use (e).

frequent in the area, the LPI for the concave sites is about three times greater than the ones obtained for the other hillslope forms. However, contributing area–total area draining to each cell (4 m^2) in the grid-proved to be the most important topographic attribute controlling the spatial distribution of landslides in the studied sites. As shown in Fig. 5c, topographic sites with high contributing areas, although having a low frequency in the basins (1–4%), present the highest LPI values. These results attest the major role played by surface and subsurface hillslope hydrology in determining slope stability in these areas and that special attention should be given to topographic hollows which control the spatial distribution of saturated zones and the development of critical pore pressures capable of triggering landslides. Therefore, these topographic hollows should be considered hazard areas and potential sites for detailed field-based investigations.

Hillslope orientation (aspect) shows a strong inheritance from bedrock structure, especially metamorphic foliation. As presented in Fig. 5d, about 70% of the hillslopes inside the two basins face SW, W and NW. Associated with this general trend, the highest LPI values are observed for hillslopes facing N and NW directions. Although hillslope orientation also indirectly affects other factors that contribute to landslides, such as precipitation, soil moisture, vegetation cover and soil thickness, it is evident that in these basins, bedrock structure plays the most important role in controlling hillslope orientation.

Fig. 5e shows the results associated with vegetation cover and land use. Although degraded forest is the most frequent vegetation cover in the basins, LPI values are relatively similar for grasslands, degraded forest and preserved forest. The highest LPI value shown by preserved forest should be analyzed with caution because this vegetation type is spatially associated with the steepest hillslopes. This suggests that other investigations, based on statistical analysis, should be carried out in order to characterize better the relationships between these parameters.

4.2. Landslide susceptibility analysis

A variety of scenarios were simulated and landslide susceptibility estimated using the model SHALTAB. As described before, the spatial distribution of the topographic attributes were obtained from the detailed digital elevation model generated for the two basins. Soil parameters were considered constant inside the two basins and were obtained from field evidences and from previous studies carried out in the area. Fig. 6 shows the results for one of these scenarios, derived from applying Eq. (1) to each cell (4 m^2) in the grid, describing landslide susceptibility in terms of log Q_c/T values. The classes vary from unconditionally stable (orange) to unconditionally unstable (gray), with a continuous gradation between these two end-points (Fig. 6a).

The model results are evaluated by comparing the predicted unstable sites with the actual location of landslide scars in the basins, triggered by the intense rainstorms of February 1996. It can be observed that some sites considered unconditionally unstable by the model did not fail in the 1996 event. Field investigations attest, however, that these areas are constituted by bedrock escarpments. In this case, therefore, field observations are in agreement with the model prediction that soil cannot be sustained on these steep hillslopes.

Fig. 6b shows a detailed view for part of the studied area, defined by the box selected in Fig. 6a. The spatial distribution of landslide scars is overlaid in this figure, allowing a general comparison between the estimated landslide susceptibility and the actual landslide location for the 1996 event. The results attest a general agreement between the predicted most unstable classes (gray and red) and the spatial distribution of landslides in the basins. Besides, the elongated geometry of these most unstable classes, as predicted by the model SHALSTAB, is similar to those of the landslide scars mapped in the field. Fig. 7 shows SHALSTAB efficiency in estimating landslide susceptibility in these two basins. For a total number of 92 landslide scars mapped in the 1996 event, 95% was predicted by the two most unstable classes of the model (Guimarães et al., 1999; 2003; Guimarães, 2000). Other studies are being carried out in the area in order to

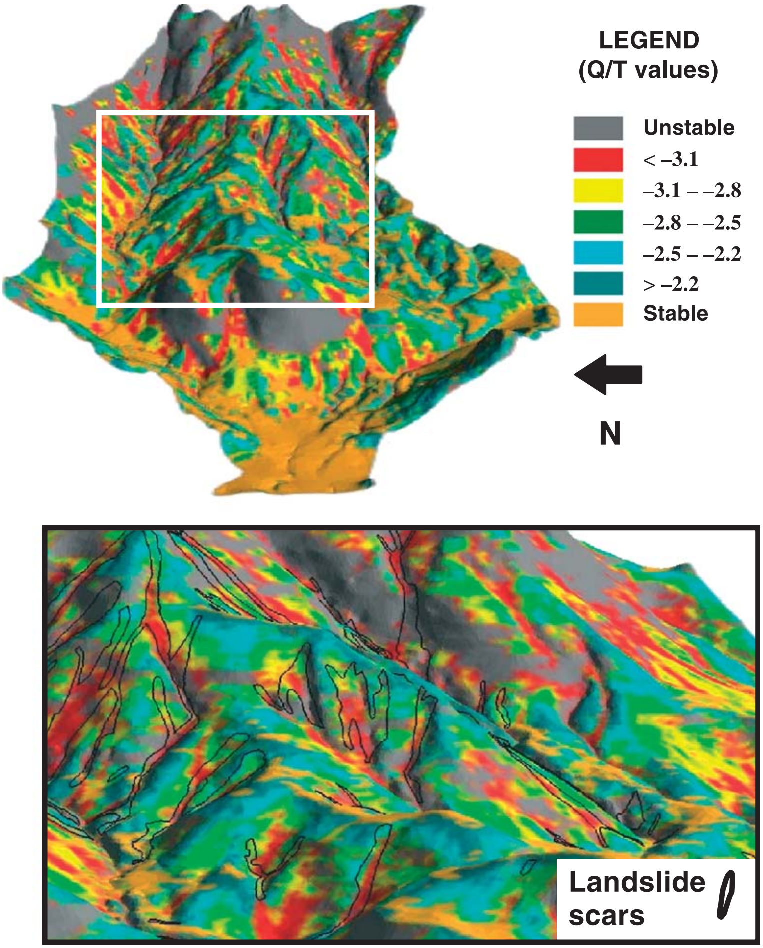

Fig. 6. a (top) shows landslide susceptibility (in a 3D-view) as predicted by SHALSTAB, expressed in terms of log Q_c/T, for one the simulated scenarios ($C/z=2$; $\phi=45°$; $\rho_s=1.5$ g/cm^3) in the two studied basins: Quitite (left) and Papagaio (right). The figure at the bottom (b) shows approximately a detailed view of the area inside the square in a. It can be noticed a good agreement between landslide scars from 1996 and the most unstable sites predicted by the model. The area on the far right of b was not analyzed by the model.

incorporate the effects of grid scale (Gomes, 2002) and distributed soil parameters in model efficiency.

4.3. Variations in soil hydraulic conductivity

About 90 field measurements of saturated hydraulic conductivity, using the Guelph permeameter, were carried out in Papagaio basin. About 95% of the obtained values, describing the entire weathered profile to a 5.0m depth, varied from 1.0×10^{-6} to 9.0×10^{-5} m/s (Vieira and Fernandes, 2003). The general trend observed in areas with

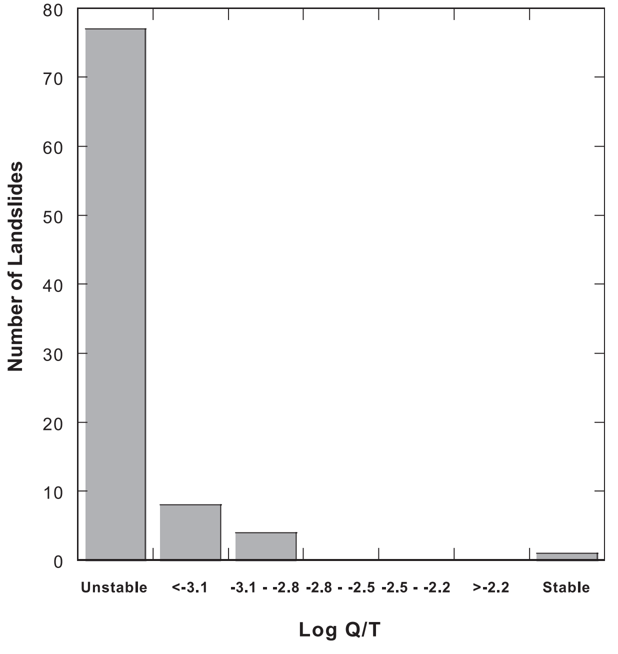

Fig. 7. Analysis of model efficiency in predicting the observed landslide in Quitite and Papagaio basins. More than 95% of the landslide mapped in the basins, resulting from the 1996 event, were inside the three most unstable classes predicted by the model.

thick soils (>3.0 m thick), is the one in which hydraulic conductivity tends to decrease from the surface with depth, attaining a minimum value at depths between 0.8 and 2.0 m, associated with the pedologic B Horizon. For greater depths, hydraulic conductivity tends to increase slowly with depth due to the presence of a sandy-silt saprolite.

In many investigated profiles, however, important variations in saturated hydraulic conductivity were observed. Frequently, inside the upper portions of the weathered profile, as the one presented in Fig. 8, hydraulic conductivity may decrease by two orders of magnitude along only 0.30 m of the profile, in this case between the depths of 0.30 m and 0.60 m. Similarly, another hydraulic discontinuity can be observed in this profile about 5.0 m depth, where conductivity decreases by about one order of magnitude along 0.60 m of the weathered profile.

The general trend described in this profile represents a broader pattern, frequently observed in the basin. Saturated hydraulic conductivity is usually more close to the surface, inside the A Horizon (0.30–0.50 m depth), decreasing sharply in the clay-rich B Horizon, generally at depths about 1.0–1.8 m. From this point on, hydraulic conductivity tends to increase with depth, attaining relatively high values inside the sandy-silt saprolite. In the lower portions of the weathered profile, closer to the bedrock, at depths varying from 2.0 to 6.0 m, hydraulic conductivity starts to decrease sharply again (Vieira, 2001; Vieira and Fernandes, 2003). Although the numerical simulations presented here with the model SHALSTAB did not consider lateral and vertical variations in the

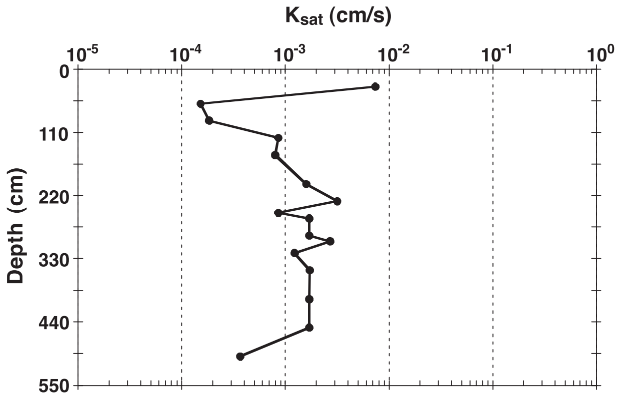

Fig. 8. Variation of saturated hydraulic conductivity (K_{sat}) with depth for a typical weathered profile of the area, until 5.0 m depth. The measurement, obtained in the field using the Guelph permeameter, show important variations in K_{sat} along short distances, attesting the potential effects of hydraulic discontinuities inside the weathered mantle.

hydraulic conductivity, new experiments are being carried out in order to incorporate such effects.

5. Final considerations

Landslides in Quitite and Papagaio basins, two typical drainage basins of the Serra do Mar in southeastern Brazil, triggered by the intense summer rainstorms of 1996, attested the important role played by topography in controlling the spatial distribution of landslides in Rio de Janeiro city. Field mapping and mathematical simulations carried out here, suggest the existence of threshold angles for landslide initiation in these areas. They also point to the fact that contributing area (per contour width) and hillslope form were the main topographic attributes defining critical conditions for landsliding. Therefore, in contrast to the commonly used procedures for landslide hazard estimation in southeastern Brazil (Fernandes et al., 2001), based mostly on the role played by slope angle, future studies should also incorporate contributing area and hillslope form. In addition, GIS-based analyses, using high-resolution digital elevation models, allow a rapid and efficient characterization of these parameters over relatively large areas.

Although the simulations carried out in this study, using the SHALSTAB model, at this point did not incorporate the spatial variability of soil properties, the results attest the model efficiency in predicting unstable sites in the Serra do Mar, Rio de Janeiro. Such behaviour supports the idea that most landslides in such a complex tropical environment are driven by topographic variables. In addition, the predicted unstable areas are not only represented as dots or squares in a map, but elongated polygons (similar to landslide scars) directly associated with the routing of water and sediments on the hillslopes. Therefore, it is suggested that a process-based model like SHALSTAB, mainly based on the topographic controls on landslides, has a high potential for predicting landslide susceptibility in steep tropical areas where soil properties are still not well known.

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