Geological and morphological study of the Jiufengershan landslide triggered by the Chi-Chi Taiwan earthquake

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Abstract

The Jiufengershan rock and soil avalanche is one of the largest landslides triggered by the Chi-Chi earthquake Taiwan 1999. The landslide destabilized the western limb of the Taanshan syncline along a weak stratigraphic layer. It involved a flatiron remnant, which was almost entirely mobilized during the earthquake. The avalanche was slowed down by NS trending ridges located downstream along the Jiutsaihu creek. The landslide affected a 60 m thick and 1.5 km long sedimentary pile composed of shales and sandstones, which dip ~22° SE toward a transverse valley. The triggering mechanism and the sliding process were analyzed by means of geological and morphological data from aerial photographs and observed in the field. A high-resolution airborne Light Detection and Ranging (LiDAR) image taken 2.5 years after the landslide allows the identification of morphological structures along the sliding surface and the landslide accumulation. The sliding surface shows several deformation structures such as fault scarps and folds. These structures are interpreted in terms of basal shear stresses created during the avalanche. Three major joint sets were identified at the sliding surface. The isopach map of the landslide was calculated from the comparison between elevation models before and after the earthquake. The coseismic volume of mobilized material and landslide deposit data are $42 \times 10^6$ m$^3$ and $50 \times 10^6$ m$^3$, respectively. The geometry of the landslide accumulation in the field has an irregular star shape. The morphology of the deposit area shows a sequence of smooth reliefs and depressions that contrast with the neighboring ridges.

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1. Introduction

The 21st September 1999 Chi-Chi Taiwan earthquake ($M_L=7.3$, $M_W=7.6$) caused severe damage...
including 2412 deaths, more than 11,000 casualties, over US$11.8 billion capital lost (4% of Taiwan’s GNP). This earthquake triggered 9272 landslides with sliding surfaces larger than 625 m² (Liao, 2000). The Jiufengershan landslide is one of the major mass wasting processes triggered by Chi-Chi earthquake. It occurred as a rock and soil avalanche that buried 39 persons.

The Jiufengershan landslide (120.84°E, 23.96°N) is located about 12 km to the north of the epicenter. Weathered, jointed rock and soil materials slid along the bedding plane creating a deep-seated rock and soil avalanche during the earthquake.

The volume of the landslide deposit comprises between 30 and 90 × 10⁶ m³, according to estimations published in literature (Kamai et al., 2000; Huang et al., 2002; Shou and Wang, 2003; Wang et al., 2003). The preliminary estimation of thickness of the displaced material is between 30 and 50 m (Huang et al., 2002). The rocks involved in the movement are Miocene in age and mainly composed of thick-bedded muddy sandstones with intercalated thin shale beds. Weathered rocks and soil were transported downslope for about 1 km. The rock and soil was avalanche deposited against several mountain slopes located downstream, infilling valley gorges and damming two small rivers.

This paper presents different approaches to describe the geological and morphological features of the Jiufengershan landslide. The geomorphological analysis is performed by means of the interpretation of aerial photos taken before and after the Chi-Chi earthquake. The topography of the landslide is studied by means of a high-resolution digital elevation model (DEM) from airborne laser altimetry data. The comparison between the topography before and after the slide allows precise calculation of the initial volume of rock and soil that was destabilized, as well as the geometry and isopach map of the landslide deposit. Field observations at different scales were performed along the sliding surface and the landslide deposit, which give new insights on the mechanical behavior of the landslide. The influence of large landslides on the erosion of the Taiwan western fold-and-thrust belt is discussed in the light of volume estimations of large landslide accumulations generated during the Chi-Chi earthquake.

2. Geological and morphological settings

Taiwan Island is located on the junction point of the Luzon Arc and the Ryukyu Arc, where the Philippine Sea Plate (PSP) converges toward the Eurasian Plate (EP) with a velocity of 8.2 cm/year in NW direction (Fig. 1) (Yu et al., 1997). Eastward of Taiwan the PSP subducts beneath the EP along the Ryukyu Trench. Southward of Taiwan the crust of the South China Sea (continental margin of the EP) subducts beneath the PSP along the Manila Trench. The crust of the Luzon arc (on the PSP) is overriding the Eurasian margin (Seno, 1977; Suppe, 1981; Seno et al., 1993; Malavieille et al., 2002).

The tectonic structure of west-central Taiwan can be subdivided into three major belts: the Western Coastal Plains, the Western Foothill Belt and the Hsuehshan Range Belt (Ho, 1975, 1976, 1986). The Western Foothill Belt is defined by an imbricated west vergent fold-and-thrust belt, bounded by the Che-lungpu and Shuilikeng faults (Fig. 1) (Suppe and Namsom, 1979; Suppe, 1980, 1981; Hung and Wiltschko, 1993). It can be divided into an Outer Foothill zone and an Inner Foothill zone, which are delimited by the Shuangtung fault. The Outer Foothill zone consists of Pliocene and Pleistocene sedimentary rocks. The Inner Foothill zone consists of late Oligocene to Miocene sedimentary rocks (Huang, 1986; Mao et al., 2002).

The Jiufengershan landslide is located in the Inner Western Foothill zone and affected middle to late Miocene sandstones with interbedded shale layers. Fig. 2 shows the geological map of the sliding area, as well as a hill-shading model obtained from a high resolution DEM (details on the DEM are given in the following section). The stratigraphic formations from bottom to top in the study area are defined as follows: Tanliaoti Shale (TL), Shihmen Formation (SM), Changhukeng Shale (CHb, CHm, CHt) and Kueichulin Formation (KC) (Huang et al., 2000, 2002; Wang et al., 2003). The Tanliaoti Shale is an early Miocene formation composed mainly of thick shale beds and subordinate interbeds or laminations, overlain by alternations of shale and siltstone beds. This formation is exposed about 800 m toward the east of the sliding area and it is bounded by sub-metamorphic rocks of the Hsuehshan Range Belt located on the footwall of the
Fig. 1. Geodynamic map of Taiwan island and enlargement of central Taiwan illustrating the regional tectonic structures in the Chi-Chi epicentral zone and the location of the Jiufengershan landslide. The ages of geological formations are indicated as follows: Q=Quaternary, P=Pliocene, M=Miocene, O=Oligocene, E=Eocene.
Shuilikeng fault. The Shihmen Formation is composed of light grey thick massive sandstones. This early Miocene formation can be divided into three members in this area. The bottom member is 145 m thick and it is composed of massive sandstones alternated or interbedded with shale layers. The middle member is 110 m thick and it is mainly composed of muddy sandstones. The top member is 165 m thick and it is composed of massive sandstones. The Shihmen Formation is more competent than the others, creating characteristic morphological structures such as the narrow river gorges and ridges. The Changhukeng Shale is an early to middle Miocene formation composed mainly of shales and sandstones. This formation is also subdivided into three members (Fig. 2) (Wu, 1986; Huang, 1986). The bottom member (CHb) is 80 m thick, and it consists of massive glauconite-rich shales with laminations or interbeds of sandstones and shales. The middle member (CHm) is 90 m thick and it is mainly composed of several thick to massive muddy sandstone layers. The top member (CHt) is 180 m thick and it consists dominantly of shales with alternations of sandy silts and shales. The sliding materials involved the top layers of the middle member (over roughly 10 m) and the overlying layers of the top member (CHt) (over roughly 40 m). The Pliocene Kueichulin Formation is exposed in the axial zone of Taanshan syncline and it is composed of calcareous massive fossiliferous sandstones. The contact between the Changhukeng Shale and the Kueichulin Formation is probably a disconformity (Tang, 1977; Huang, 1986).

The mobilized strata of the Changhukeng Shale are situated in the western limb of the Taanshan syncline. This syncline is an asymmetric fold, which plunges gently southward. The dip angle of the western limb varies between $15^\circ$ and $25^\circ$ in the nearby area. Almost all of the sliding material was removed from the western limb. The dip of the eastern limb is steeper (between $50^\circ$ and $75^\circ$). Another major structure is the Shuilikeng fault, situated about 1 km to the east of the landslide. The fault trace is characterized by Eocene–Oligocene sub-metamorphic rocks, which overthrust the Miocene strata described previously. This fault is associated with the formation of the Taanshan syncline.

A flatiron shaped remnant defined by south-east-dipping strata characterizes the morphology of the syncline’s western limb in the sliding area. This triangular-shaped landscape is limited by two notches, which confine and bound the sliding area.

The Jiutsaihu creek flows northward along the axial zone of the Taanshan syncline, and then deviates eastward showing incised meanders and gorges, as it crosses the competent Shihmen Formation. The eastward segment of the creek was partially filled by the Jiufengershan rock and soil avalanche deposit (Fig. 2). The Sezikeng creek is a tributary of the Jiutsaihu creek and it is located in the western limb of the Taanshan syncline. Its drainage basin shows a rectangular drainage network, which follows the dip of the strata and preexisting joint directions.

The morphology of the landslide zone has been analyzed by means of aerial photographs dating from 1977 to 2001. Fig. 3 illustrates two stereopair aerial photographs taken before and after the Chi-Chi earthquake. The morphology of the area did not change significantly between 1977 and 1999 before the earthquake. Nevertheless, the expansion of agriculture during this period was clearly identified from the type and density of vegetation. Before the earthquake, the sliding area was cultivated with betel palms and bamboo groves, as well as tropical fruits and vegetables. Small consequent gullies were also present along the dip slope (Fig. 3A). Note that foothill erosion is not observed along the Jiufengershan flatiron remnant. Thus, this mechanism did not contribute to the generation of the landslide. Nevertheless, chemical weathering affected the upper layers of the flatiron slope, in particular near the foothill. Before the earthquake, the sliding surface
Fig. 3. Stereopairs of aerial photographs taken before and after the Chi-Chi earthquake (explanations in the text).
located on the dip slope was bounded by three linear zones, which can be observed in the aerial photographs from 1979 (Fig. 3A). The northern boundary corresponds to an EW scarp face; the landslide mobilized the rock strata up to the free end of the scarp. The westward boundary trends N–NW and was defined by a scarp face that limited a flatiron remnant; a small creek flowed parallel to the scarp, creating a decametric notch. Infiltration of groundwater from this notch along stratigraphic and tectonic discontinuities probably increased the pore water pressure along the sliding surface. The southward boundary of the sliding surface trends NW and is oriented downslope. This boundary corresponds to a small V-shaped valley created by a small consequent creek. The sliding surface is located roughly beneath the notches created by the incision of two creeks along the boundaries described above. Thus, three linear zones may be identified with free boundaries that do not contribute to the shear strength of the slope.

The aerial photographs taken three days after the earthquake showed water seepage and small springs, located near folded and faulted rocks observed on the sliding surface. These observations suggest that groundwater was flowing along interlayered permeable rocks and preexisting joints. The water table was probably located well above the sliding plane. Nevertheless, no precipitation was recorded during a period of several days preceding the earthquake, which occurred during the rainy season. The shear strength of the rocks was probably at its lowest annual level, as a result of high pore pressures in the zone of saturation. Thus, the landslide triggering by seismic shaking was enhanced by high pore water pressure at the décollement level.

The sliding direction along the dip slope is indicated by large striations and grooves observed on the sliding surface, which trend SE (Fig. 3B). The rock and soil avalanche was slowed down by morphological obstacles such as the NS trending ridges observed downstream along the Jiutsaihu creek. These ridges are situated in the eastern limb of the Taanshan syncline and are composed of competent sandstones from the Shihmen Formation. The avalanche deposit dammed the Jiutsaihu and Sezikeng creeks, creating two small lakes located southward of the deposit (Fig. 2).

3. Geomorphic and geological analysis of the landslide area

The high-resolution DEM from airborne LiDAR (Lillesand and Kiefer, 1994; Priestnall et al., 2000) was taken in April 2002 (2.5 years after the Chi-Chi earthquake). The precision for this type of data set is up to 1 m horizontal and 0.15 m vertical (Shih and Peng, 2002). LiDAR data was used to generate a 1 m × 1 m surface grid DEM, which shows highly detailed geomorphic and geological features around the study area (Fig. 4). The main scarps and structures of the rock and soil avalanche are well identified in the LiDAR image.

3.1. Sliding surface

The sliding area has been encircled by a dashed curve in Fig. 4 and it mainly involves two superposed flatiron remnants (shown in Fig. 3A). The uppermost flatiron was almost entirely mobilized during the earthquake and transported by the avalanche.

Fig. 5 shows a precise contour level map calculated from the LiDAR data set. The sliding area shows a regular slope, which increases slightly downslope. The northern boundary of the sliding area is limited by the flatiron crest. The E–W trending scarp face of the flatiron showed small debris that slid northward to the valley bottom. Small post-seismic slumps showing circular crowns were mapped on the sliding surface, close to the flatiron crest. A disturbed zone located well above the bottom of the valley is observed to the N–E of the sliding surface. This zone shows debris accumulation deposited by the avalanche and incipient block slides (Fig. 6A).

Two scarps are observed along the western boundary of the sliding area (Fig. 4): the main scarp is aligned along the small creek described in the aerial photos, and it marks the contact between the uppermost flatiron located eastward and less disturbed rock material located westward. This rock material shows metric to decametric scarps dipping eastward and downslope, and they result from small translational slides and slumps that were initiated during the earthquake. In the southwestern lateral flank, we observed a major lateral scarp, trending S–E. The scarp, which is about 30 m high, is connected with several en-echelon transtensional cracks. These cracks
are associated with tensional stresses generated by the drag of the avalanche.

Fig. 7 shows a block diagram of the sliding area, using the LiDAR data set. The sliding surface was initially covered by a thin accumulation of debris. These deposits have been progressively washed out, clearing the sliding surface. Linear accumulations oriented downslope are still observed along the slope. The Longnan path crosses the sliding surface at mid-altitude and is roughly parallel to the strike of the dip.

Fig. 4. (A) LiDAR hill-shading image showing detailed morphological features of the landslide. The sliding area (SA) is contoured by a dashed line; the deposit area (DA) is contoured by a dashed and dotted line. LP=Longnan path. (B) Stereographic projection of the main directions of preexisting joints and faults observed in the upslope and downslope area of the sliding surface. The sliding surface (defined by the stratification S₀), and the striations generated by the rock and soil avalanche, are also plotted.
slope (Figs. 3, 4 and 7). Hereafter, the sliding surface is divided into an upslope and a downslope surface, which are located above and below this country path, respectively. The sliding surface is affected by several metric to decametric faults and folds. The orientations of the principal fault and joint families observed in the upslope surface are plotted in the stereograms (Fig. 4B). The average strike and dip of the bedding plane in the upslope surface is N36°8'E/21°8'S (small arrows indicate the striations created during the avalanche). The main orientations of preexisting joints and faults are N42°8'W/85°'S, N46°E/60°'N and N82°E/75°'N. The average bedding plane in the downslope surface is slightly higher (N36°8'E/23°8'S). The main orientations of preexisting joints and faults are quite similar: N32°W/74°S, N42°E/58°'N and N86°E/75°'N.

The upslope surface shows a linear structure corresponding to a monoclinal fold (Fig. 7). This structure is parallel to the bedding strike and it crosses the entire sliding surface for more than 800 m. Its orientation is coherent with one of the joint sets. Fig. 8A illustrates several cross-sections oriented perpendicularly to the fold, showing a small inflexion of the sliding surface. The monocline fold was possibly created by the shear drag of the rock material transported during the avalanche (Fig. 9A). Intense shearing oriented downslope may have activated secondary décollement levels located along weak shale layers, below the sliding surface. Movement along the décollements is absorbed by folding and by small thrusts observed at the toe of the inflexion zone. These structures activate weak zones such as preex-
isting joints and faults. Some of the joints exhibited horizontal slickenfibers of calcite-rich minerals, associated with early tectonic deformation. The joints were filled with sheared mud injected from nearby saturated clay layers. These structures result from liquefaction of unstable layers during the rock and soil avalanche.

Fig. 6. (A) Disturbed zone located above the Longnan path to the N–E of the sliding surface. Block rotation and incipient sliding are extensively observed. (B) Detail of an open joint located at the sliding surface, showing sheared mud injected from nearby saturated clay layers. These structures result from liquefaction of unstable layers during the rock and soil avalanche.

observed: (a) tension cracks oriented perpendicularly and parallel to the fold axis, (b) conjugated shear faults and (c) reverse faults. The density of fractures is greater near the toe of the fold, where joint orientation is more chaotic.

Fig. 9B illustrates a different hypothesis to explain the surface inflexion: the fold is associated with the activation of a secondary décollement level located beneath the sliding surface. The sliding surface shows a step, passing from a higher stratigraphic level upslope, to a lower stratigraphic level downslope. This mechanism seems less realistic because it would have created extensional structures in the step zone.

The sliding surface shows a gentle convex curvature between the monoclinal fold and the Longnan path. In this area, several preexisting joints cut through the in-situ rock layers and propagate through the overlying landslide deposits, indicating post-seismic sliding.

The central zone of the downslope sliding surface exhibits several linear fault scarps characterized by small flexure zones (Figs. 4, 7 and 8B). These flexure zones resulted from the downslope shearing of the near surface layers during the avalanche. Differential movements along the décollements are mostly absorbed by compressional faults located beneath the flexure zones (Fig. 10A). A few normal faults and open joints resulting from local extensional stresses in between the blocks were also observed. Both reverse and normal faults tend to reactivate preexisting master joints and faults. Preexisting faults show dextral en echelon fractures, calcite slickenfibers and mud-rich fault breccias composed of chaotic sandstone blocks.

The northeastern zone of the downslope sliding surface shows a sequence of shallow folds (four anticlines and three synclines) (Figs. 4, 7 and 10B). Fig. 8C illustrates a topographic section that cuts across the folded layers. The top layer is strongly fractured and it has been partially eroded. Tension cracks are observed mostly in the axial zone of the folded strata. Anticline folds have metric amplitudes and asymmetrical limbs, indicating an upslope deformation vergence. This vergence direction is opposite to the downslope shear sense generated by the avalanche. These folds resulted from compressional and shear stresses applied on these layers during the dynamic phase of the avalanche (Fig. 9C). Compressional stresses may have induced buckling of the top
layers, which are buttressed downslope. Downslope shearing can activate décollement levels located along the underlying weak layers. The fold axes are oriented NNE and are slightly oblique to the strike of the sliding surface. This orientation may have resulted from the 3D boundary conditions of the sliding process: folded strata are also buttressed laterally toward the NE, at the periphery of the sliding area. The southwestern zone of the downslope sliding surface shows open cracks along preexisting conjugated fracture zones (Figs. 4 and 10C). The fractures separate poorly deformed block of metric to decametric scale. This chocolate tablet structure is related to local tensional stresses during the sliding process.

3.2. Landslide deposit

The avalanche debris was transported through a transverse valley, which flows across several mountain crests trending NS (Figs. 3 and 4). The morphology of the deposit area shows a sequence of smooth reliefs and depressions that contrast with the neighboring steep ridges (Fig. 5). After the earthquake, an elongated depression was observed at the toe of the sliding surface. This small basin was filled progressively with debris transported by small fans from the sliding surface, as shown in the hill-shading image taken 2.5 years after the earthquake (Fig. 4). The rock and soil avalanche created the two larger lakes observed on the southward boundary of the deposit. These lakes as well as the former depression have been drained by means of artificial channels, which are easily recognized in the LiDAR image. The channels were dredged in order to prevent subsequent geologic hazards such as debris flows. The flat morphologies located near the channels are artificial and they do not correspond to subsequent flood deposits. The structure and the composition of the debris deposit were observed in fresh outcrops located in the channel banks and at the surface. The deposit shows a chaotic mixture of small rock fragments and jointed blocks, ranging in size from a few centimeters to more than 20 m. Rock fragments are enclosed by a matrix of weathered shales. The nature and structure of the deposit suggests that the rock and soil avalanche began as a translational slide consisting of large blocks that were rapidly broken up to form an avalanche, as the material was pulverized in transit. The blocks probably split and broke apart along preexisting fracture zones that are well identified beneath the sliding surface (Fig. 4).
4B). Flow conditions were favored by the high sliding velocities and the moisture from the ground water zone.

4. Volume estimation

To illustrate the morphological evolution of the landscape associated to the landslide, we compared the relief before and after the earthquake. Fig. 11A shows an isopach map illustrating the difference in height between the LiDAR image (taken in 2002) and a DEM of the topography in 1986. Strictly speaking, this map illustrates the superposition between coseismic and post-seismic erosion processes. Post-seismic material transport between 1999 and 2002 is minor in the sliding zone, in spite of the intense typhoons that affected central Taiwan during this period. Post-seismic erosion and deposition were mainly observed in artificial drainage channels and in the small lakes formed by the debris deposit, respectively. The 1986 DEM has been baseline corrected around the landslide zone in order to minimize the vertical discrepancy between the two elevation models. For this, the mean altitudes of several stable areas were calculated for both DEMs. The vertical offsets were quite the same for all areas, with differences of less than 1–2 m in between them. The 1986 DEM was shifted vertically in order to match the 2002 DEM. The data set of the 1986 DEM...
is defined by a 40 m square grid, and small topographic features are frequently smoothed. The differences in height between the two baseline corrected DEMs around the sliding area are generally metric and have a minor effect on volume estimations. The isopach lines are shown at 10 m intervals: negative and positive values indicate net accumulation or depletion of rock material, respectively. The average thickness of the sliding strata is around 60 m. The isopach values around 40 m are located in areas where thick debris deposits overlay the sliding surface (e.g. near the Longnan path). The thickness of the destabilized geologic layers is greater than the estimation published by Huang et al. (2002), which ranges between 30 and 50 m. The isopach map indicates that the translational slide affected a homogeneous sedimentary sequence, extending over a large portion of the flatiron structure. In cross section A–A′, we have plotted two topographic profiles across the landslide, calculated from the LiDAR image and from the 1986 DEM (Fig. 11B). The mobilized strata, indicated in light grey and grey, extended for 1.5 km along the sliding surface and the junction area. The ratio between the thickness and the length of the sliding layers is quite low (0.04). The main décollement level has been prolonged beneath the debris deposit area, considering the sliding surface and the initial depth of the transverse valley. A dashed line indicates the unexposed sliding surface. The estimated volume of rock and soil materials that slid during the avalanche is roughly $42 \times 10^6$ m$^3$. The volume of post-seismic eroded material (between 1999 and 2002) in the sliding area is $\sim 1 \times 10^6$ m$^3$. Our volume estimation is different from other values published in literature based on simplified geometrical analyses. The following estimations have been reported: 30 $\times 10^6$ m$^3$ (Huang et al., 2002), 35 $\times 10^6$ m$^3$ (Shou and Wang, 2003), 50 $\times 10^6$ m$^3$ (Wang et al., 2003) and 90 $\times 10^6$ m$^3$ (Kamai et al., 2000).

The length of the landslide accumulation reaches 1.2 km along the transverse valley. The landslide accumulation is mainly located within a kilometric scale depression that was filled by the rock and soil avalanche. The geometry of the landslide accumulation in the field has an irregular star shape, which follows roughly a topographic contour line. The lakes are located on the upstream valleys that converge toward the deposit. The average thickness of the deposit is

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Fig. 10. Detailed views of morphological structures observed in the downslope area of the sliding surface. (A) Fault scarp showing fractured and tilted stratigraphic layers. (B) Jointed and fractured folded layers. (C) Open cracks observed along preexisting conjugated fractures zones, which separate poorly deformed blocks.
between 60 and 80 m, reaching 110 m above the transverse valley. This estimation is much lower than the values published by Huang et al. (2002), who proposes that the maximum debris thickness reaches 175 m based on a preliminary topographic analysis.

The isopach lines are deviated along the artificial channel. The estimated volume of the landslide accumulation is roughly $46 \times 10^6$ m$^3$. The debris volume calculated from the LiDAR image is underestimated, because a large amount of sediments have
been transported downstream between 1999 and 2002. Debris weathering and compaction after the avalanche also reduces the volume of deposit. The co-seismic volume of the landslide accumulation is ~12% larger than our estimation. This volume is slightly greater than the initial volume of rock and soil materials. Volume changes depend on several processes such as the decompaction of sediments during the avalanche, increasing the volume in 19%.

5. Landslides and erosion

The Jiufengershan avalanche is one of the largest landslides triggered by the Chi-Chi earthquake. To visualize the size of the slide, we may compare it with a homogeneous layer of 1.4 mm thickness, extending all through Taiwan Island. This value is comparable with yearly uplift rates registered in the Taiwan orogenic belt. Likewise, two other kilometric scale areas showing extremely large mass-wasting processes were also observed near the epicentral zone. These zones, located respectively in the Tsaoling monocline and the Jiujiufon massif, mobilized huge volumes of materials from the hillslope to the foot of the slope. The Tsaoling rock and soil avalanche mobilized $125 \times 10^6$ m$^3$ of weak rocks and soil material along a monocline slope dipping $\approx 14^\circ$. The landslide deposit dammed a river, creating a lake that was drained by an artificial channel. The Jiujiufon (99 peaks) massif composed of poorly cemented conglomerates was denuded over a surface of ~12 km$^2$, generating large accumulations of cobbles that were deposited in a large network of stream channels. Narrow valleys that incise very steep hillslopes characterize the morphology of the massif. The volume of material mobilized by the earthquake in this massif was roughly $20 \times 10^6$ m$^3$.

The total volume of sediments transported downhill by these three landslides is ~0.2 km$^3$. The landslide accumulations deposited in the stream channels, delivering new material to be transported by the rivers during annual typhoons that are frequent in Taiwan (Dadson et al., 2003). Dadson et al. (2004) argue that most of the landslides triggered by the Chi-Chi earthquake remained confined to hillslopes and only 8% delivered sediment to rivers. However, large landslides can deliver huge amounts of sediment to the channel network even if they seldom happen. Rock and soil avalanche accumulations would have probably been mobilized by subsequent debris flows if the lakes located upstream had not been drained by artificial channels. The volume of these landslide deposits is comparable with the annual volume of sediments transported by the fluvial network to the ocean, in Taiwan. Dadson et al. (2003) have estimated that the average annual sediment yield is 500 Mt·year$^{-1}$, which is roughly equivalent to 0.25 km$^3$ (taking an average bulk density of ~2.0 t/m$^3$). These results suggest that large landslides have had a major impact in the past erosion rates in the Taiwan western thrust belt.

6. Conclusions

The Jiufengershan landslide affected a 60 m thick and 1.5 km long flatiron remnant of shales and sandstones, dipping $22^\circ$ toward a transverse valley. The flatiron was bounded by scarp faces and creeks (Fig. 12), which correspond to free limits that do not contribute to the shear strength of the slope. The sliding area shows several fault scarps and folds resulting from the downslope shearing of the near surface layers during the avalanche.

The avalanche began as a translational slide consisting of large blocks of rocks and soil that were rapidly broken up to form a flow, as the material was pulverized in transit. Fluidization along the sliding surface created an excess pore pressure, which reduced the shear strength (Vardoulakis, 2002). The landslide deposit is located within a kilometric scale depression that was filled by the rock and soil avalanche. The average thickness of the deposit is between 60 and 80 m. It consists of a chaotic mixture of small rock fragments and jointed blocks.

The Chi-Chi earthquake generated a few kilometric scale landslides that transported materials from the hillslope to the foot of the slope. Rock and soil avalanches affecting weak sedimentary rocks in gently dipping fold limbs largely contribute to the erosion of active mountain belts such as the Taiwan western fold-and-thrust belt. In these situations, the erosion signal from the incision of the rivers at the base of the landslide to the hillcrest may propagate very rapidly.

Future studies should assess the initiation and propagation mechanisms of the landslide, using precise morphological and structural data as well as earthquake strong motion records and rock mechan-
ical behavior. Post-seismic sediment transport at the sliding surface and the debris deposit may also be studied in order to analyze the erosion processes.

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