Suppl. Geogr. Fis. Dinam. Quat.
III, T. 3 (1999), 115-135

FOURTH INTERNATIONAL CONFERENCE ON GEOMORPHOLOGY - Italy 1997
Session: Volcanic Geomorphology

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RESEARCH PROBLEMS AND POTENTIAL CONTRIBUTIONS TO RELATED EARTH SCIENCES: AN OVERVIEW

INTRODUCTION

Not only volcanism has a considerable geomorphological significance, but Volcanic Geomorphology is original for at least two reasons. (1) By contrast to ordinary landforms, volcanic landforms are the results of *opposing constructive and destructive* forces; hence, volcanic landforms are first-order objects to study processes of landform growth and erosion. (2) In contrast to ancient landforms carved in basement rocks, volcanoes typically have an *ephemeral* life; thus, volcanic constructs enable us to measure and compare rates of growth and denudation, because erosion acts rapidly as soon as volcanoes start growing.

We present an assessment of research problems and potential contributions which stems from a review of recently published papers in volcanic geomorphology and volcanology. This represents a challenge because volcanology and geomorphological tools (DEM, GIS, airborne imagery) have expanded at an unprecedented rate over the past 20 years. Our review emphasizes two major needs. First, data should be integrated from ground-based observations, statistical analyses, laboratory experiments, and numerical models, as increasing numbers of studies attest to the close relationship and complex interplay of eruptive and depositional processes, characters of deposits, and environmental factors. Second, efforts to measure rates of geomorphic processes acting at all scales on volcanoes and volcanic landscapes, would help to drive the shift towards process-oriented geomorphology (Thouret, 1992).

1. SIGNIFICANCE OF VOLCANIC GEOMORPHOLOGY

The significance of Volcanic Geomorphology can be amplified through (a) selected criteria on which determination of volcano and related landforms can be based, (b) improvement of the classification of volcanic landforms through quantitative investigation, which blends morphometry and studies based on ground observations, remote sensing data, and laboratory experiments, and (c) the diversified use of airborne images and digital data acquired through radar and satellites, and combined with DEM's data, to facilitate the morphological analysis of volcanoes.

1.1. *Improvement of classification of volcanoes and related landforms*

Classical classifications of volcanic landforms are based on types of activity, magmas, and erupted products (e.g., Cotton, 1944; McDonald, 1972). Improved classifications should also be based on geomorphic scale, constructional versus erosional origin, mono- versus polygenesis, types of activity, and type and volume of magma and erupted material. Hereafter, we distinguish six main types of volcanic constructs and erosional landforms (e.g., Ollier, 1988; Francis, 1993). However, many rapidly constructed volcanic landforms are not volcanoes at all, such as flood basalt continental or submarine plateaus, and ignimbrite sheets from large calderas.

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1.1.1. Classification of volcanoes and related landforms

a) Monogenetic landforms and fields:
- cinder or scoria cones, (Surtseyan) tuff cones, and (Taalian) tuff rings,
- maars (subaqueous and subaerial) and diatremes,
- intra- or subglacial volcanoes: tuyas (table mountains) and mobergs,
- endogenous and exogenous domes,
- lava flows and fields, including small-scale lava-flow forms,
- continental flood basalts and plain basalt provinces,
- ash flows and ignimbrite sheets, plains, and plateaus.

b) Polygenetic volcanoes and calderas:
- stratovolcanoes: simple with summit crater, composite with sector collapse scar and/or a caldera; compound or multiple volcanoes,
- intermediate-silicic multivent centres that lack a central cone; rhyolithic centres; silicic volcanic lava field with multiple domes and calderas,
- calderas types: explosion (somma), collapse-explosion (Krakatoa), collapse on Hawaiian shield volcano, collapse in basement and resurgent caldera (Valles), large and complex resurgent calderas (Toba),
- volcano-tectonic depressions termed «inverse volcanoes» (Taupo Volcanic Zone).

c) Shield volcanoes:
- Hawaiian shields and domes; Galapagos, Icelandic, and scutulum-type shields.

d) Volcanic landforms resulting from eruptive and/or erosional processes:
- avalanche caldera from a flank failure of magmatic, gravitational, or mixed origin,
- erosional calderas (e.g., Haleakala, Maui; La Réunion cirques).

e) Volcanic landforms resulting from denudation and inversion of relief:
- eroded cone; eroded pyroclastic-flow deposit and sheet,
- inverted small-scale forms: necks, culots, dykes,
- eroded lava flow, inverted relief and planeze (e.g., Pain & Ollier, 1995),
- roots of paleo-volcano, cauldron, and hypovolcanic complex.

f) Morphological changes in volcanic-surrounding landscapes:
- volcano construct and induced change in drainage pattern at a regional scale
- drainage blockage, avulsion, impoundment and lake-breakout, etc.

1.2. Why do we need to improve existing classifications?

1.2.1. Several examples illustrate the pitfalls of classical classifications

The classification of tuff cones and tuff rings, for example, has been based on effects of explosive magma/water interactions on morphology (e.g., Cas & Wright, 1987): tuff rings and tuff cones are thought to result from relatively dry and wet eruptions, respectively, which are related to low and high mixing ratios of water to magma. Sohn (1996) has challenged this prevailing model for tuff rings and tuff cones in South Corea, arguing that the morphological variations are directly caused by depositional processes (pyroclastic-surge dominated in tuff rings and fallout-dominated in tuff cones), irrespective of water/magma mixing ratios. The depositional processes are interpreted to be in turn controlled by a number of fundamental controls, which include depositional settings, type, level, and lithology of aquifers, strength of country rocks, ground-water behavior, and properties and behavior of magma. These controls determine the explosion depth, conduit geometry, mode of magma-water interaction, magnitude of explosion, eruption-column behavior, and subsequent depositional processes.

1.2.2. Existing classifications face increasing complexity in landform generation

Several examples illustrate the coexistence of several eruption styles and eruptive sequences within the lifetime of a complex volcano, and the contribution of several mechanisms to caldera formation. For instance, giant tuff cones and calderas were described in arc volcanoes, such as the Ambrym caldera and tuff cone, Vanuatu (Robin & *alii*, 1993). Previously considered as an effusive basaltic volcano, Ambrym consists of a basal shield volcano topped by an exceptionnally large tuff cone surrounding a 12-km-wide summit caldera. The tuff cone may be considered as mainly basaltic. Interpretation of the tuff series implies intervention of external water and suggests both explosive and collapse mechanisms for this type of non-classical model of caldera formation at a basaltic volcano.

In addition, we need to improve the classification to understand magmatic systems and achieve hazard assessments of the activity of a young volcano. One example stems from the study of Cerro Negro volcano, Nicaragua: is this a young composite stratocone or a scoria cone (McKnight & Williams, 1997)? This question bears on the fact that hazard assessment of the current activity of Cerro Negro depends heavily on the morphological type of volcano. Criteria of age and size that are usually used for this assessment are not adequate for active, young volcanoes. Other criteria on which that determination can be based are magma production rates, cone morphology, and eruption style.

1.3. How can we improve existing classifications?

Detailed research has been undertaken over the recent years on a few volcanic landforms and on the detailed morphology of lava flows, in terms of morphometry, comparative morphology, and processes, based on remote sensing, ground-based observations, and laboratory experiments.

1.3.1. Recent results acquired on large-scale landforms

Shield volcanoes, for example, have been the focus of recent morphological studies on size, distribution, and magma output rate (Mexican shield volcanoes: Hasenaka, 1994), and on morphology and mechanism of eruptions (Icelandic shield volcanoes: Rossi, 1996). Additional results acquired on smaller landforms such as dome origin and behavior illustrate the evolution from a descriptive approach (Scarth, 1995) to a semi-quantitative approach (Blake, 1990).

1.3.2. Detailed morphology of lava flows

Recent studies on lava flows have been based on three approaches: fieldwork and morphometry, remote sensing, and laboratory experiments.

a) Ground observations and quantitative, comparative morphology have beeen carried out on icelandic lava flows and small-scale landforms on lava flows (Rossi, 1997; Rossi & Gudmunsson, 1996).

b) Remote sensing helps to outline lava flows and detail lava-flow morphology using spaceborne Radar images (TOPSAR, SIR-C radar), satellite images (Landsat TM, SPOT), and photographic data. Digital multispectral data such as the thermal infrared multispectral scanner (TIMS) images help to infer the chemical and physical properties of the surface materials and to map lava flows. For example, the Mauna Loa lava flows were mapped in great detail (Kahle & *alii*, 1995; Kauahikau & *alii*, 1995) using NASA'S Thermal Infrared Multispectral Scanner (TIMS) and Space Shuttle radar SIR. Lava flows can be followed up and mapped using data obtained from the spaceborne advanced very high resolution radiometer (AVHRR), whose quantitative analysis allows estimation of active lava area, thermal flux, effusion rates, and total flow field volume. Estimates of eruption rate and total flow field volume of the 1991-1993 Mount Etna effusive eruption (Harris & *alii*, 1997) are in agreement with published ground-based estimates of 5.8 m^3/s and 235 x 10^6 m^3 (Calvari & *alii*, 1994; Tanguy & *alii*, 1996). Stevens & *alii* (1997) carried out a more accurate estimate of the 1991-93 lava-flow volume (231±29 million m^3), using EDM-based field survey of the surface of the lava-flow field and one panchromatic SPOT image. The results were digitised, interpolated and converted into a DEM, constructed from a 1:25,000 contour map of the area. Digital elevation data from TOPSAR, an airborne synthetic aperture radar system that uses interferometry to derive topography, were used by Rowland (1996) to determine slope distributions, proportions of lava fows and vents, and lava flow thicknesses and volumes on Fernandina Volcano, Galapagos Islands. The concentration of vents on the summit platform, five to eight times as much as on coastal plain and apron, supports previously proposed mechanisms for producing higher elevations and steeper slopes in the central part of the volcano.

c) Laboratory simulations and comparative morphology have been carried out to reproduce morphologies observed on the sea floor. Using laboratory simulations, Gregg & Fink (1995) were able to reproduce submarine flow morphologies commonly observed on the sea floor, typically classified as pillowed, lobate, or sheet flows, while monitoring the physical conditions under which they form. Four submarine lava-flow morphologies are considered to be diagnostic of specific effusion rates: jumbled, folded and lineated sheets, and striated pillows.

1.3.3. A wealth of volcanic landforms on ocean floors

Recent investigations confirm that the ocean floors, in particular the mid-ocean ridges, are home to over 60% of the Earth's volcanoes and concentrate the essential part of the present and past volcanic activity. Investigations of oceanic ridges have emphasized the fast-spreading East Pacific Ridge, the medium-spreading Juan de Fuca ridge, the slow-spreading Mid-Atlantic Ridge, and the super slow-spreading SW Indian Ridge. Volcanic constructs include axial topographic highs, abyssal hills, and seamount populations which show a spatial density and characteristic height in accordance with the spreading rate (Smith & Cann, 1992; Mendel & Sauter, 1997).

a) Based on swath bathymetric coverage combined with high-resolution side-scan images, the three-dimensional perspective view of the axis of the slow-spreading Mid-Atlantic Ridge (Smith & *alii*, 1997), shows volcanic constructs and faults similar in size and shape to those observed at subaerial rift zones such as Hawaï and Iceland. The overall shape of the axial zone is that of a major graben composed of an inner valley floor and bordered by valley walls along normal faults. The inner valley floor is the primary site of crustal construction, and most segments contain large axial volcanic ridges within their valley floors that are the principal sites of lava extrusion: seamounts, hummocks, fissures, and smooth flows. Axial volcanic ridges range in size up to several hundreds of meters high, 1.5 km wide, and several to tens of kilometers long. Small ($50<h<300$ m) near-circular seamounts are distributed over the valley floors.

b) The topographic features known as abyssal hills (typically 10-20 km long, 2-5 km wide, 50-300 m high, and oriented approximately perpendicular to the spreading direction) characterize > 30% of the ocean floor (Macdonald & *alii*, 1996), being the most abundant geomorphic structures on Earth. Submersible-based investigations show that Pacific abyssal hills are created on the East Pacific Rise as horsts and grabens which lengthen with time. Hills are bounded on one side by ridge-facing scarps produced by normal faulting, and on the other by more gentle slopes produced by volcanic growth faulting.

Seamounts (Smith & Cann, 1997; EOS, 1996), guyots (Smoot & King, 1993; Smoot, 1995), and shoaling volcanoes (McPhie, 1995) play a significant role in crustal construction and in constructional-erosional processes. Seamounts play a significant role in crustal construction at the mid-oceanic ridges, at least in the slow-spreading ridges such as MAR. Spreading segments contain a prominent axial volcanic ridge. Ridges are composed of piled up seamounts and hummocky flows, and are interpreted as the primary sites of crustal construction. Small magma pockets with slow eruption rate produce seamounts; small magma bodies with somewhat higher eruption rates produce hummocky fissure-fed flows.

1.4. Geodynamic and tectonic settings of volcanic constructs

The passage of magma through the crust and lithosphere is controlled by crustal lithospheric stress-field and local stress-field configurations, and resultant fractures, i.e., normal faulting, thrust faulting, and strike-slip faulting (Cas & Wright, 1987).

1.4.1. Tectonic effects on volcano and caldera location, morphology, and formation

To evidence the relationships between volcanic complexes or caldera location, morphology, and tectonics, three approaches have been undertaken, on a morphotectonic basis (e.g., statistical analysis of the geometry of drainage patterns and stream directions), on structural and remote sensing data (e.g., shape and formation of large calderas), and more recently, on laboratory experiments and modelling.

a) Classical morphological studies include the statistical analysis of the geometry of drainage patterns and stream directions controlled by tectonics and the identification of morphotectonic features. For example, the main morphological characters of volcanic complexes of Latium in Italy (Caputo & *alii*, 1989; Buonasorte & *alii*, 1991; Trigilla, ed., 1995) are strongly controlled by four prevailing tectonic directions for Vulsini, and three for the Sabatini and Colli Albani areas. These studies allow to infer where and how the structural setting of the sedimentary units and the recent tectonic activity of the area «control» the location and shape of the calderas.

The tectonic influence as a structural frame on the shape and the formation of large caldera complexes has been inferred from remote sensing and ground-based structural analysis. Based on 2D-images produced by draping SPOT satellite images over a DEM, the relationships between the geodynamic setting, the regional faults, and the calderas of Toba and Tondano (Indonesia) are described in terms of evolution in a pull-apart basin (Lecuyer & *alii*, 1997). The links between the main morphologic and structural features on the submerged portions of volcanic edifices such as Panarea, Stromboli in the Aeolian Islands (Gabbianelli & *alii*, 1993), have been carried out through electroacoustic and high resolution seismic profiles in the SE Tyrrhenian sea. Both complexes show a preferential development along NE-SW lineaments, which coincides with the regional structural trend of this sector of the Aeolian structure. Faulting, caldera collapse, and tectonic tilting were interrelated and fundamentally influenced by activation of the NE-SW fractures.

1.4.2. Volcanic constructs can influence their tectonic setting and magmatic system

a) Volcano building on thin oceanic lithosphere causes the lithosphere to sag downward into the asthenosphere. As much as a half or two-thirds of the upbuilding of the volcanoes of Hawaï may be offset by lithospheric subsidence (Lipman, 1995; Peterson & Moore, 1987).

b) The tectonic effect of volcanic constructs on faulting has been tested through laboratory experiments and modelling. The effect of volcanic construct on rift fault patterns (van Wyk de Vries & Merle, 1996) is exemplified in three cases: Fieale volcano (Asal Rift, Djibouti), Axial and Brown Bear seamounts (Juan de Fuca Ridge), and Maderas volcano (Nicaragua). Analogue models show that increased fault throw as the volcano is approached is caused by an interaction of the regional stress field with that set up by the volcano mass. For faults to be reoriented there must be a ductile layer below the volcano (hot crust at mid-ocean ridge, weak sedimentary strata, etc.). Increased volcano mass and size and lower brittle/ductile ratios lead to increased fault curvature. Volcanoes on one side of a rift may capture it, forming the axis of a new rift. By concentrating extension, magma is more easily erupted. A positive feedback between increased extension and magma eruption rate will lead to rift narrowing, which can favor the formation of oceanic crust (van Wyk de Vries & Merle, 1996).

c) Conversely, volcanic constructs and gradual volcano spreading influence faulting and rifting (Merle & Borgia, 1996; van Wyk de Vries & Merle, 1996), as well as the slope instability of the edifice (van Wyk de Wries & Francis, 1997). A volcano of sufficient size induces stresses that may deform its substratum. In turn, this deformation feeds back stresses which deform the edifice. Both stresses and deformation influence the evolution of magma by varying the boundary conditions of magmatic systems.

2. CONTRIBUTION TO GEOLOGY AND VOLCANOLOGY:
Volcano growth and destruction, large-scale instability, and relationships with tectonics, sedimentation, and drainage

To contribute to geology and volcanology, volcanic geomorphology should (a) produce detailed geologic-geomorphological maps and expand the use of accurate chronological frameworks based on high resolution techniques, and collect compositional data through eruptive sequences, to identify eruptive or constructional «stages» in complex volcanoes, (b) expand the capability of geomorphology in landscape history reconstruction, thus contributing to the understanding of processes of building and destruction of volcanic edifices through geological and historical times, and (c) analyze the structural factors which contribute to the catastrophic collapse of volcanoes.

2.1. Growth and destruction-denudation of volcanoes

Growth and destruction of volcanoes are the result of the complex interplay of endogenous and exogenous processes. One of the major tasks devoted to volcanic geomorphology is to reconstruct the volcanic landscape and landform history, in order to (1) unravel the building and destruction stages and processes, and (2) calculate long- and short-term growth and denudation rates over time, in response to eruptive style, tectonic uplift or spreading, climate, and sea-level change.

2.1.1. Rapid processes and rates of growth and destruction of volcanic constructs

Unlike ordinary mountains, which are formed by slow uplift and erosion, volcanoes are *constructed rapidly.*

a) Growth and mature stages of oceanic shield volcanoes

Seven stages characterize the evolution of the oceanic shield volcanoes located above an active hot spot, such as the Hawaïan volcanoes (Peterson & Moore, 1987; Decker & *alii*, 1987; Moore & Clague, 1992; Tilling & Dvorak, 1993; Rhodes & Lockwood, 1995): (1) initial stage; (2) shield-building stage including three submarine, sea-level, and subaerial substages; (3) capping stage; (4) erosional stage; (5) renewed volcanism stage; (6) atoll stage; and (7) late seamount stage. The seven volcanoes comprising the island of Hawaï and its submarine base are, in order of growth, Mahukona, Kohala, Mauna Kea, Hualalai, Mauna Loa, Kilauea, and the still submarine volcano, Loihi. The first four have completed their shield-building stage. The island of Hawaï has grown at an average rate of about 0.02 km^2/yr for the past 600 kyr and presently is close to its maximum size (Moore & Clague, 1992). On each volcano, the transition from eruption of tholeiitic to alkalic lava occurs near the end of shield building. The rate of southeastern progression of the end of shield building, and hence the postulated movement rate of the Pacific plate over the Hawaïan hotspot, in the interval from Haleakala to Hualalai is about 13 cm/yr. On the basis of this rate and an average spacing of volcanoes of 40-60 km, the volcano requires about 600 thousands years to grow from the ocean floor to the time of the end of shield building. They arrive at the ocean surface about midway through this period (Moore & Clague, 1992).

b) Caution needed to reconstruct the long-term evolution of stratovolcanoes

Recent detailed studies of time-volume-composition concerning the long-term behavior of stratovolcanoes in active arcs (e.g. Mount Adams, Cascades: Hildreth & Lanphere, 1994; Tatara-San Pedro complex, Chile: Singer & *alii*, 1997) challenge previous studies in stating that *caution should apply to reconstruction of the life history of so-called polygenetic volcanoes.*

Subdivision of complex stratovolcanoes into eruptive and constructional «stages» needs detailed geologic mapping, accurate high-resolution geochronology, and compositional data. Stratovolcanoes commonly grow in spurts: construction of an imposing cone needs take only 1-5 % of the active lifetime of the volcano, i.e. construction rate of 1-5 km^3/ky. Higher rates are exceptionnal. Discussion of volumetric eruptive rates can be misleading without adequate time and space scales (Hildreth & Lanphere, 1994).

Stratovolcanoes can remain active between the widely spaced episodes of peak productivity, as much as for half a million years. Documented examples of greater longevity are rare. «Dormancy» is an anthropocentric notion and generally only an upper-crustal condition, without fundamental change in deep-level magmatic processes (Hildreth & Lanphere, 1994).

Stratovolcanoes need not develop large upper-crustal magma chambers and need never evolve towards a caldera-forming stage. Arc calderas that result from collapse of shallow reservoirs beneath stratovolcanoes (Mazama, Krakatau, Santorini) are usually associated with large eruptions of rhyodacitic to rhyolitic ejecta. Recurrent eruptions of small batches of dacite at irregular intervals and its secular alternation with varied andesite and even andesitic basalt shows that there is no standard sequence, no unidirectional progression, and certainly nothing predetermined in the evolution of stratovolcanoes (Hildreth & Lanphere, 1994).

Andesite-dacite production in the focal region and coeval basaltic activity on the periphery have coexisted at several documented stratocones (e.g., Mount Adams: Hildreth & Lanphere, 1994). Scarcity or abundance of surrounding mafic cinder cones and lavas has nothing to do with maturity of the stratovolcano system. The term *parasitic* should be abandoned as it implies dependence upon the main stratocone and promotes the view that peripheral mafic eruptions are leaks from a central chamber or conduit. Virtually the opposite is true: andesitic stratovolcanoes are the derivative features and flank failures are lateral breakouts from a central conduit system, whereas peripheral basalts, having their own conduit from mantle or deep-crustal depths, are the more fundamental (Hildreth & Lanphere, 1994).

2.1.2. Volcano growth and erosion result from complex and interdependent processes

a) Erosion-prone glacial periods have altered the fast growth record of mid-latitude stratovolcanoes.

The life history of Andean volcanoes provides an important insight into how we may interpret the chronological record of glacially eroded volcanic edifices. The Tatara-San Pedro complex in the Northern Chilean Andes (Singer & *alii*, 1997) preserves about 55 km^3 of lavas that erupted during seven eruptive sequences from at least three central vent regions. Remnant, unconformity-bound sequences of lavas are separated by lacunae that represent significant periods of erosion. Estimated growth rates for the two young volcanoes are 0.2 to 0.3 km^3/ky, i.e., three to five times greater than a growth rate estimated from all preserved lavas in the complex (0.06 km^3/ky). Removal of up to 50-95 % of the material erupted between 930 and 200 ka by repeated glacial advances largely explain this discrepancy, and it raises the possibility that episodic erosion of mid-latitude frontal arc complexes may be extensive and common. Hence, Singer & *alii* (1997) raise fundamental questions bearing on the interpretation of the life history of glacially eroded stratovolcanoes.

Frontal arc volcanic complexes (of the Andes) may remain active for about 1 my, during which time many changes in vent position, magma composition, and magma reservoir processes can occur. Thus, several earlier volcanoes of possibly comparable dimensions can be largely obliterated.

Preserved lavas and erosional hiatuses in dissected volcanoes, consistent with global records of terrestrial ice advances, offer additional valuable source of information bearing on the timing and extent of glaciations, to bracketing the age and extent of tills, drifts, and moraines in regions of high elevation or maritime climate.

Accumulation of ice volumes leading up to glacial maxima is a slow process requiring 80-150 ky. By contrast, major deglaciations proceed rapidly (< 15 ky) to glacial minima. In contrast to selectively eroded young units and summit portions of any volcanic edifice, the oldest lava at the base of a sequence overlying an erosional unconformity may be emplaced very shortly after the glacial maximum. Thus, the correspondence in timing between lacunae in stratigraphic sequences and the global ice-volume maxima recorded in the astronomical time scale cannot be considered merely fortuitous (Singer & *alii*, 1997).

b) Repetitive and instantaneous mass-wasting destruction characterize the development of oceanic volcanoes.

In the Canarian islands with Quaternary volcanic activity, the construction has been closely controlled by rift-type volcano-tectonic features, characterized by a tight cluster of recent emission centres piled up along narrow dorsal ridges. Two main types of rifts can be defined in terms of geometry: simple, as in La Palma, or triple, as in Tenerife and Hierro. Carracedo (1994) provided a hotspot-based schematic model for the genesis of a complex «Mercedes»-type stellate rift zone on one of the Canary Islands. The geometry of the complex Canarian three-branched rifts separated by angles of 120° suggests a least-effort fracture as a result of magma-induced vertical upwards loading. The concentration of the recent eruptive activity and the depressions that may have generated by gravitational slides are also included in the model (see discussion by Marti & *alii*, 1996). The importance of this model is that it may help to explain the main landforms and the two sources of volcanic hazards on the Canary Islands (Carracedo, 1996a,b): (1) the concentration of recent (Quaternary-Holocene) eruptive vents, and therefore, the statistically most probable location of future eruptions, and (2) the genesis of the main open-towards-the sea Canarian depressions and, subsequently, the areas where catastrophic slope failures may be pending.

In the Canaries, two main types of large depressions are open toward the sea: straight-walled and arcuate head basins such as the Orotava and Guimar valleys in Tenerife and crescent-shaped coastal embayments like the El Golfo in El Hierro. Their genesis has been the subject of a long debate. Faulting and collapse in giant gravitational landslides (Carracedo, 1996a,b; Masson, 1996) are now preferred to a mainly erosional origin. An alternative hypothesis of lateral expansion of the Guima and Orotava valleys based on multiple cycles of erosion and filling of valleys by lava flows has also been proposed (Palacios, 1994). However, recent detailed mapping of areas of the submarine flanks of the Canaries has revealed much new evidence of seven major landslides over the past 500,000 years, strengthening the case for large-scale slope failure as a principal agent of island destruction (Masson, 1996).

2.2. *Large-scale instability processes, erosional landforms, and debris-avalanches*

As a consequence of rapid construction, many volcanoes are liable to massive flank or slope failures resulting from structural instability. Slope failures produce extremely mobile debris avalanches that can travel long distances beyond the flanks of volcanoes at high velocities. More than 20 major slope failures have occured globally during the past 500 years, a rate exceeding that of caldera collapse (Siebert, 1996). The main characteristics and origins of flank failures and related deposits were described by Voight & *alii* (1983), Siebert & *alii* (1987), Crandell (1988), Glicken (1991), Moore & *alii* (1994), McGuire & *alii* (1996), as follows.

Massive landslides create specific morphology and deposits, i.e., horseshoe-shaped rentrants into the edifice, and a high, steep-sided break-away scarp having a amphitheater shape. Debris avalanches typically form a hummocky terrain with water-filled depressions and steep flow margins, and thick hummocky deposits with block and matrix facies of largely unsorted and unstratified angular-to-subangular debris.

A relationship exists between avalanche runout and failure volume. The maximum failure volume of subaerial volcanoes typically does not exceed 10% of the edifice volume. The ratio of vertical drop H to travel length L ranges from 0.09 to 0.18 (av 0.13) for Quaternary volcanic avalanches between 0.1 and 1 km^3 in volume and from 0.04 to 0.13 (av 0.09) for avalanches > 1 km^3 (Siebert & *alii*, 1987). The ratio of H to L for volcanic avalanches is much lower than the ratio for non-volcanic deposits of similar volume, suggesting that low-rigidity, perhaps partially fluidized avalanches are capable of travelling great distances. Exceptional runout distance travelled by avalanches over 100 km have been reported at stratovolcanoes such as Nevado del Colima in Mexico (Stoopes & Sheridan, 1992).

The widespread occurrence in a variety of tectonic settings suggests that slope failure may be the dominant catastrophic edifice-modifying process. Steep-sided andesitic and dacitic stratovolcanoes, with a relief that can attain several km and upper slopes that can exceed 30°, are obvious candidates for slope failure (e.g., Socompa volcano, Northern Chile: Wadge & *alii*, 1995). Steep-sided but less voluminous lava-dome complexes are also particularly susceptible to slope failure. The air of permanence of large, low-angle shield volcanoes belies their inherent instability. Particularly noteworthy 70 landslides have occurred on the Hawaiian ridge, where they have removed volcano-flank sectors that exceed 1000 km^3 in volume (Moore & *alii*, 1989, 1994; Iverson, 1995). A series of maps from several sources document the importance of landslides, best exposed in the submarine realm, in the growth and decline of Mauna Loa and adjacent volcanoes (Moore & Chadwick, 1995). Catastrophic slope failures are neither rare nor unique in the lifetime of a volcano. The summit edifice of Augustine, Alaska, has repeatedly collapsed and regenerated, averaging 150-200 years per cycle, during the past 2,000 years (Beget & Kienle, 1992). The unprecedented frequency of summit edifice failure was made possible by sustained lava effusion rates over 10 times greater than is typical of plate-margin volcanoes.

The origins of flank failure are bound to three types of events (Siebert & *alii*, 1987; Siebert, 1996): magmatic eruption of Bezymianny type (1956), non-magmatic explosions of Bandai type (1988), and cold avalanches of Ontake type (1784). Several structural and geomorphic factors contribute to flank failure (Voight & *alii*, 1983; Vallance & *alii*, 1995; McGuire & *alii*, 1996): (1) steep dip slopes with alternating competent lavas and unconsolidated pyroclastic materials, (2) zones of weakness within the upper volcanic edifice, owing to hydrothermally altered rocks converted to clay minerals of low yield strength, accompanied by the boiling of supercritical hydrothermal fluids (Lopez & Williams, 1993), and (3) local extension promoted by parallel dike swarms (Hausback & Swanson, 1990).

Yet, important but enigmatic questions remain unsolved. What processes can trigger slope instability at low-angle shield volcanoes? Forces in addition to gravitation must trigger the landslides (Iverson, 1995). Tilling & Dvorak (1993) invoke a seaward displacement of the southern flank of Kilauea volcano, while Owen & *alii* (1995) measured its rapid deformation to be as much as 10 cm per year, based on GPS measurements. The observations can be explained by slip on a low-angle fault beneath the south flank combined with dilatation deep within Kilauea's rift system, both at rates of 10-15 cm per year.

What fundamental structural causes might lead to an edifice collapse? Van Wyk de Vries & Francis (1997), diverging from previous works which emphasized differences in eruption style associated with flank failure (Siebert & *alii*, 1987), argue that the volcanic edifice itself can contribute to the weakness of its bedrock. In contrast to radially spreading volcanoes, preferential spreading in one direction is critical to collapse development; whereas radial spreading tends to generate inward-dipping graben which inhibit collapse, sector spreading generates failure-prone outward-dipping structures. Spreading in a preferential direction may be caused by buttressing, by the regional slope of basement beds, by regional stress, by weak basement or by high fluid pressures under one side (van Wyk de Wries & Francis, 1997).

2.3. *Relationships between volcanism, tectonics, sedimentation, and drainage*

Volcano growth and erosion, as well as tectonics, bear influence on sedimentation and the evolution of drainage pattern in volcanic landscapes.

2.3.1. Modification of drainage pattern

In his study of the morphotectonic development of southeast Australia, Ollier (1995) shows several examples of drainage modifications in catchments affected by the formation and erosion of huge volcanoes and by the retreat of the Great Escarpement. Large volcanoes on or near the Great Divide resulted in the formation and superimposition of radial drainage in the vecinity of the volcano, and major drainage disruption in neighbouring regions, even affecting big rivers.

2.3.2. Complex relationships between volcanism, tectonics, and sedimentation

The geomorphic evolution of the Sunda volcanic complex and the Bandung area, a large intramontane basin surrounded by volcanic highlands in Java (Dam & *alii*, 1996; Nossin & *alii*, 1996), illustrates complex landform-determining processes, such as tectonic subsidence, paroxysmal eruptions, volcanism-induced faulting/rifting, drainage system adaptations, and intramontane lacustrine sedimentation. Geomorphological and sedimentological studies reveal that the morphology of the Bandung basin and the Sunda-Tangkuban Perahu volcanic complex encompassed seven phases during the Middle-Late Quaternary, in particular since 125,000 years B.P.: (1) the early Bandung basin during Middle Quaternary, (2) the start of lacustrine sedimentation and the formation of an enclosed intramontane basin ca. 125,000 yr B.P., (3) paroxysmal volcanic eruptions and formation of the Sunda caldera and the east Lembang fault ca. 105,000 yr B.P., (4) ongoing lacustrine sedimentation 105,000-50,000 yr B.P., (5) second phase of Plinian and caldera-forming eruptions, and Bandung volcaniclastic fan development 50,000-35,000 yr B.P., (6) high lake levels and lacustrine sedimentation, and formation of the west Lembang fault 35,000-20,000 yr B.P., (7) small eruptions of Tangkuban Perahu volcano and minor basin subsidence and sedimentation, and minor fluvial erosion after 16,000 yr B.P. The events mark the significance of the Sunda-Tangkuban Perahu volcanic centre during the Late Quaternary; the Sunda volcano collapsed into a caldera in which later the Tangkuban Perahu volcano developed. Moreover, these eruptions controlled regional sedimentation and determined landform development in the great basin area. In the vecinity of the eruption centre, volcano-tectonic faulting formed the conspicuous E-W Lembang fault that controlled distribution of volcaniclastic sediments and the initiation of a new drainage system in the Lembang area.

3. CONTRIBUTION TO PHYSICAL VOLCANOLOGY: interplay of construction and denudation processes throughout eruptive activity

Volcanic geomorphology can contribute to physical volcanology through the (a) observation and assessment of topographic effects on transport, erosion, and deposition of volcanogenic flows, in particular high-velocity pyroclastic currents, (b) analysis of the complex relationships between eruption phenomenology and eruptive processes, aiming to quantify opposing constructive and destructive forces, and (c) identification of the factors which govern the emplacement of volcaniclastic deposits, to assist in the determination of sources and climatic/tectonic conditions for volcaniclastic sedimentation.

3.1. Complex interplay of eruptive activity with geomorphic processes

3.1.1. Topographical effects on transport and emplacement of pyroclastic flows

Observations of recent or ongoing eruptions suggest that high energy relief may exert effects on transport and deposition of primary pyroclastic deposits, such as the flow-surge laid down by the blast on 18 May 1980 at Mount St. Helens (Fisher, 1990, 1995), and the block-and-ash flows at Unzen (1991-1993). The decoupling of pyroclastic currents is accentuated by encounters with steep mountain ridges of high relief (Fisher, 1995). In regions of rugged topography, the height of barrier ridges, slopes angles and gradients of the ground surface greatly influence the effectiveness of decoupling processes and flow directions. The topographical effect of breaks in slope on the flanks of stratocones contributes to the decoupling of two zones within a pyroclastic flow (e.g., at Merapi, 1994): the channelized, dense, gravity-driven part of the pyroclastic flow and the unconfined, dilute and low-concentration, turbulent part of the flow whose destructive effects project far beyond the valley channels. The interactions of pyroclastic currents with topography include blocking, downslope drainage, formation of secondary pyroclastic flows in valleys and development of dividing streamlines and decoupling (Fisher, 1995).

3.1.2. Erosion during waxing phases of pyroclastic flows-surges

Waxing phases of blasts and pyroclastic flows can exert erosion on landforms such as erosional furrows formed during the 1980 eruptions of Mount St. Helens (Kieffer & Sturtevant, 1988; Kieffer, 1995). Field estimate for the erosion rate by the 18 May 1980 lateral blast is 20.6 $kg/m^2/s$, corresponding to an erosion depth of nearly 1 m in a time of 30 seconds. The 7 August 1980 pyroclastic flow triggered an erosion rate of 14.3 $kg/m^2/s$, corresponding to 2 m of erosion in 5 minutes (Kieffer & Simonds, 1995). Pyroclastic flows generated in the 19-20 April 1993 eruption of Lascar Volcano, Chile, produced spectacular erosion features (Sparks & *alii*, 1997). Exposed bedrock and boulders suffered severe abrasion, producing smoothed surface on coarse breccias and striations and percussion marks on bedrock and large boulders. Erosional furrows developed with wavelengths of 0.5-2 m and depth of 0.1-0.3 m. Erosive features were produced where flows accelerated through topographic restrictions or where they moved over steep slopes. Much of the erosive phenomena are attributed to lithic clasts which segregated to the base of the flows, from the lithic-rich part of the pyroclastic flows. The erosive features, distribution of lithic clasts and deposit morphology indicate that the 1993 flows were highly concentrated avalanches dominated by particle interactions.

3.1.3. Monitored and measured processes of dome growth and magma supply

Processes of dome growth can be observed through aerial photographs and satellite images, and, in some cases, measured through a real-time monitoring network (e.g., Mount St. Helens, Santiaguito: Anderson & *alii*, 1995; Unzen, Merapi, Monserrat). A dacite dome at Unzen volcano, for example, grew in two pulses, mainly exogenously when it was small and the effusion rate was high (4 x 10^5 m^3/day; May 1991 to Februrary 1993), but endogenously when the dome became large and the effusion rate declined (5 x 10^4 m^3/day; February 1993 to August 1994). The volume of magma erupted during each pulse was 1.3 x 10^8 m^3 and 0.6 x 10^8 m^3, respectively (Nakada & *alii*, 1995).

Magma supply rates are high on recently grown domes (Redoubt, 1989-1990: Miller, 1994; Unzen, 1991-1994: Nakada & *alii*, 1995) or present-day growing domes (Soufriere Hills, Monserrat, 1996-97). At Monserrat, a dome appeared in mid-November 1995 in the English's Crater of Soufriere Hills within 4 months of the eruption's onset (MVO Team, Young & *alii*, 1997). The dome has grown endogenously and exogenously like at Unzen, while dome collapses on the NE flank have triggered many block-and-ash pyroclastic flows that were channeled in the Tar river valley towards the East. Since October 1996, a new dome has filled in the explosion vent and pyroclastic flows have resumed, even towards the West where the capital city, Plymouth, only 4 km of the crater, is therefore evacuated. Since January 1997, magma supply increased again to 2 x 10^5 m^3/day up to as much as 7 x 10^5 m^3/day (MVO Team, Young & *alii*, 1997). In February 1997, the volume of the dome was of 40 x 10^6 m^3.

3.2. Interactions in construction and denudation processes

3.2.1. Lava flow and coastal processes

Lava flows entering seawater create a prograding lava delta on the Southern coast of Hawaï island (Mattox & Mangan, 1997). For nearly 14 years, pahoehoe flows from Kilauea advance down the flank of the volcano from the Pu'u'O'o lava cone and form a system of tubes that transport lava to the coastline. An average volume of 350,000 m^3/day of lava was fed through the tube system between

1986 and 1994. During this time, 2 km^2 of new land was added to the island. Two types of interactions were observed: open mixing of lava and seawater resulting from the complete collapse of lava bench severs active lava tube; confined mixing conditions where partial collapse of lava bench submerges and fractures a portion of active lava tube (Mattox & Mangan, 1997).

3.2.2. Eruptive activity, sea-level changes, and erosion at seamounts, guyots, and shoaling volcanoes

Effects of sea-level change induced erosion have been observed at Palinuro, Italy (Gabbianelli & Colantoni, 1997) and effects of submarine eruptions were discovered in 1996 at Loihi seamount, the next-to-be born Hawaïan island, SE of Kilauea (EOS, 1996). Lithofacies associations at shoaling or active subaerial island volcanoes can be related to stages, early and late, of emergence (McPhie, 1995; McPhie & *alii*, 1993). Facies associations are sensitive to proximity to source vents and to water depth of eruption and emplacement. Lithofacies associations bear on the variation in eruption and fragmentation processes with respect to environment, especially water depth. Dry eruptions are limited to subaerial and very shallow water settings, and may be explosive or effusive. Autoclastic fragmentation operates universally although abundant hyaloclastite is restricted to subaqueous environments. Pillows are a hallmark of subaqueous lava eruption or flow of subaerial lava into subaqueous settings.

3.2.3. Interaction of collapse and exhumation processes at caldera walls

Present-day caldera walls often present a complex assemblage of cliff surfaces of different ages, which can result from repeated collapses that exhumed earlier caldera cliffs and unconformities on the same caldera rims. Geomorphological mapping at Santorini (Druitt & Francaviglia, 1992) shows that the caldera wall in the north, NE, and east preserves evidence for three generations of cliff surface, those of Minoan age ca 3.6 ka (= 17th century BC) and two earlier generations of caldera: the Skaros caldera about 76-54 ka and the Cape Riva caldera ca. 21 ka, which exhumes cliffs of the Skaros caldera. Therefore, field relationships are critical in unravelling the geomorphology of Bronze-Age Santorini immediately before the Minoan eruption, with a large diameter, a central island, and a probably flooded caldera, because the 21 ka caldera walls extend to present-day sea level at several locations.

3.2.4. Volcano-glacier interactions

Ice-clad or snow-covered active volcanoes are home to eruptions during which combined eruptive, glacial, and geomorphic processes lead to generation of primary and secondary sediment-water flows termed lahars or volcanic debris flows (Major & Newhall, 1989; Pierson & *alii*, 1990; Thouret, 1990, 1993). Five historic eruptions at four snowclad volcanoes (Tokachi-Dake, Nevado del Ruiz, Cotopaxi, and Mount St. Helens) have demonstrated that snowmelt-generated volcanic debris flows can: (1) have peak discharges as large as 10^5 m^3/s, (2) attain velocities as high as 20-40 m/s, (3) mobilize as much as 10^8 m^3 of debris, and (4) travel more than 100 km (as debris flows) in valleys draining the volcanoes (Pierson, 1995). The risk to human life from such large debris flows was tragically demonstrated in 1985 at Nevado del Ruiz volcano in Colombia, where snowmelt-triggered lahars in three of the volcano's major drainage systems killed more than 23,000 people (Pierson & *alii*, 1990; Thouret, 1990, 1993).

Attention has been drawn upon new types of volcaniclastic sediments and flows, such as the «volcanic mixed avalanches» from the November 13, 1985 eruption of Nevado del Ruiz volcano (Pierson & Janda, 1994; Vandemeulebrouck & *alii*, 1993), and unusual «ice diamicts» (comprising clasts of glacier ice and subordinate rock debris in a matrix of ice, snow grains, coarse ash, and frozen pore water) emplaced during the December 15, 1989 eruption of Redoubt volcano, Alaska (Waitt & *alii*, 1994), and the 1992 eruption of Mount Spurr, Alaska (Waitt, 1995). These transient, mixed avalanches of tephra, snow, and ice transformed to initial «snow slurry» lahars, then large dilute and small concentrated lahars in the Whangaehu catchment of Ruapehu, in 1995 (Cronin & *alii*, 1997). Volcanic mixed avalanches or hybrid wet flows (between conventional pyroclastic flows and conventional lahars) are hazardous insofar as their reduced internal friction projects destructive flows downvalley beyond the reach of dry pyroclastic currents. Such deposits at snowclad volcanoes are broad and geomorphically distinct, but they soon become extensively reworked and hard to recognize in the geologic record.

Yet, the processes involved in volcano-glacier interactions remain not well understood. Geomorphological and glaciological perturbations on ice-clad active volcanoes record a variety of processes, including rapid melting, snow and ice avalanching, surficial abrasion, and mechanical scouring or gullying (Thouret & *alii*, 1995). The loss of large volumes of snow and ice during eruptions results mainly from (1) the passage of pyroclastic flows and surges or hot blasts on the glacier, (2) the contact of subaerial lava flows or tephra with ice or snow, and (3) the eruptive or geothermal activity which melts the base of ice caps. Large volumes of meltwater released in a short time span (e.g., 38.5-44 x 10^6 m^3 in 20-90 minutes at Nevado del Ruiz on 13 November 1985) imply a high melting rate and a vigorous heat transfer from hot eruptive products to snow and ice. Preliminary melting scenarios based on vigorous deposition of hot debris on snow point to a melting rate as high as 2 cm/minute. Mechanical entrainment and comminution of snow and ice are important processes in releasing large volumes of meltwater that contribute to trigger lahars.

4. CONTRIBUTION TO SEDIMENTOLOGY: significance of volcaniclastic sediments, flows, and facies models

Volcanic geomorphology can contribute to (a) identify sedimentary facies associations and facies models for dynamic volcano-sedimentary systems, which encompass edifices, piedmonts, and both nonmarine and marine basins, (b) establish criteria for recognizing volcaniclastic deposits in old volcanic successions, and infer the role of climatic-tectonic effects on transport and deposition, and (c) analyze the characteristics of sediment gravity flows to assist in the determination of relevant parameters for modelling their behaviour.

4.1. *Volcaniclastic sediments and flows*

The rapid growth and denudation of volcanic landforms combine to cause serious erosion in drainage systems and supply huge volumes of sediments through the catchments toward neighbouring lowlands. Because tectonism and volcanism are closely associated at plate margins, and in some instances within plates, knowledge about volcaniclastic sediments and rocks may be critical for tectonics, depositional and erosional processes, as well as for understanding the economic significance of mineralized and alteration zones in both ancient and modern successions.

The importance of volcaniclastic sediments raises at least four questions: (1) How do we identify volcanogenic sediments and lithofacies associations? (2) What are the types, transport processes, and behaviour of volcanogenic flows? (3) How much and how rapidly do rates of sedimentation/erosion fluctuate? (4) How quickly do disturbed catchments recover following large eruptions?

4.1.1. Distribution and significance of volcaniclastic sediments and facies models

Sedimentary processes exert a great influence on modern volcanoes, while reworked volcanic rocks are volumetrically important and must be significant in the geological record.

Generalised facies models are based on the identification and distribution of proximal, medial, and distal nonmarine volcaniclastic facies and sedimentary cycles triggered by large eruptions around stratovolcanoes, such as Fuego, Guatemala (Vessell & Davies, 1981, *in:* Cas & Wright, 1987). Facies are features of a sedimentary unit portraying the processes of origin and source, and environment of deposition. Among all modern volcanoes, stratovolcanoes are very prone to mass-wastage because they are high topographic features and they host great volumes of easily removed fragmental material. Their growth is therefore reflected almost instantly in the sedimentary record of surrounding regions.

Composite volcanoes are complex dynamic volcano-sedimentary systems. Hackett & Houghton (1989) proposed a facies model for the composite Ruapehu volcano (New Zealand) of Quaternary age, with moderate discharge rates of magma and a wet temperate climate, and whose products are subject to rapid erosion and reworking. Long intervals are characterised by small but frequent eruptions (1943, 1995-96), providing a continual supply of debris to the surrounding ring plain. Larger explosive eruptions trigger major pulses of ring plain sedimentation. The volcano can be divided into two parts: a composite cone of 110 km^3 in volume, surrounded by an equally voluminous ring plain. Cone-forming sequences are dominated by sheet- and autobrecciated-lava flows, which seldom reach the ring plain. The ring plain is built predominantly from the products of explosive volcanism, both the distal primary pyroclastic deposits and the reworked material eroded from the cone. Much of the material entering the ring plain is transported by lahars, either generated directly by eruptions or triggered by the high intensity rain storms which characterize the region. Ring plain debris are reworked rapidly by concentrated and hyperconcentrated streams in pulses of rapid aggradation immediately following eruptions and more gradually in the longer intervals between eruptions.

4.1.2. Significance of continental volcanogenic sedimentation

The complex variations in the distribution of volcaniclastic sediments have several sources. For example, the complex distribution of the volcaniclastic deposits in an active rift can be caused by faulting, sub-basin development, sources of primary materials, and local depositional environments (Mathisen & McPherson, 1991). However, in his study on continental deposition of an extensional basin in SE Arizona, Smith (1994) offers an evaluation of continental sedimentation in response to climate, rather than tectonics. Covariance of climatic conditions (recorded in the pedogenic-carbonate isotope data) and sedimentologic parameters (sedimentation rate, channel geometry, and facies abundance) strongly suggests that climate can produce temporal variations in sedimentological processes that have heretofore been attributed only to tectonics. Hence, caution should apply in asserting tectonics as the only significant long-time period influence on depositional style in nonmarine basins. There is also a need to better understand the climate effects on hillslope processes and drainage basin evolution that regulate sediment supply and fluvial processes that deliver sediment to depositional sites, especially over time intervals of 10^5-10^6 yr.

Tephro-chronology and associated dating methods can provide a framework for volcanism, tectonism, and paleoenvironmental reconstruction of basins surrounding a volcanic province. Tephro-chronology was used as a tool in the 2-My-long geomorphic history of ignimbrite plateaus in New Zealand (Kennedy, 1994; Alloway & *alii*, 1995; Shane & *alii*, 1995). Fifty-four tephra beds span the interval 2-0.6 Ma and provide an event frequency of 1/19 ka (much higher than the ignimbrite frequency 1/100 ka). The tephra beds provide a framework for a paleoenvironmental reconstruction of the Taupo Volcanic Zone (southern North Island). Volcaniclastic transport routes from the TVZ to basins in the south and southeast, and through the site of present mountain ranges, supplied material to a terrestrial lowland fore-arc area in the interval 1.64-0.7 Ma. Uplift and deformation since 0.7 Ma have disrupted paleodrainage routes, diverting them to the north and southwest (Shane & *alii*, 1995).

4.2. *Type, characteristics, and origins of volcanogenic flows*

Volcanogenic flows can be divided into pyroclastic (primary), volcaniclastic (secondary), and epiclastic flows that erode, transport, and redeposit fragmental sediments on and around volcanoes (e.g., Cas & Wright, 1987). Among volcanogenic flows, the debris avalanches and lahars or volcanic debris flows have drawn the attention of volcanologists, geomorphologists, hydrologists, and sedimentologists since the 1980 eruption of Mount St. Helens. Many lahar flows have been watched, filmed, and monitored, especially on active stratovolcanoes (Mount St. Helens: Janda & *alii*, 1981; Sakurajima, Unzen, etc.), and considerable progress has been achieved in the 90's in understanding the processes, behaviour and rheology of lahar flows. After pyroclastic flows, lahars are the most deadly volcanic phenomena on active and dormant volcanoes, projecting effects far beyond the areas affected by the pyroclastic flows.

4.2.1. A spectrum of sediment gravity flows

A simple but synthetic classification of mass movements and flows on natural steep slopes has been proposed by Coussot & Meunier (1996) as a function of solid fraction and material type. Debris flows occupy a field between landslides and hyperconcentrated flows and encompass mudflows and granular flows on the base of cohesion and one or two-phase flow. Sediment gravity flows differ from flood flows, based on greater velocities, greater impact forces, depositional record, and longer-term effects. Debris flows transform into hyperconcentrated flows and streamflows and, with debris avalanches, are part of a spectrum of subaerial sediment gravity flows, whose main characteristics are sediment concentration, deformation rate and velocity (Pierson & Costa, 1987; Scott, 1988; Smith & Lowe, 1991).

4.2.2. Origins and characteristics of cohesive and noncohesive debris flows

Following the pioneer study by Neall (1976), recent research incorporates post-1980 trends in recognition of origins related to landslides, surges of meltwater from hot volcanic products on snow and ice, failures of natural dams formed by volcanic flows, especially debris avalanches, and glacial outburst flows (Scott & Sheridan, 1997). The dichotomy of cohesive (muddy) and noncohesive (granular) lahars is illustrated through grain-size plots (Scott, 1988; Scott & *alii*, 1995). The textures and origins of cohesive and noncohesive lahars are analysed in terms of transformation of noncohesive lahars to hyperconcentrated flows and streamflows, textural changes within transformations, transition facies, and cause of dearth of fine sediment (clay and silt). The distinction has been made between the probable syneruptivity of non-cohesive, granular flows (e.g., from pyroclastic flows) versus the possible non-eruptive origins of cohesive lahars, e.g., from debris avalanche, such as the 3.8 km^3 Osceola mudflow from Mount Rainier (Scott & *alii*, 1995; Vallance & Scott, 1997).

4.3. *Transport processes and behavior of sediment gravity flows*

Interpretations of transport processes and behaviour of debris flows (Scott & Sheridan, 1997) are based on (1) observations on modern flows and on deposits of both ancient and modern flows, (2) grain-size analysis, texture and sedimentological characteristics (critical diameter, phases and facies of debris-flow deposits, clast support versus matrix support), (3) sedimentary structures such as (a) grading, i.e. inverse grading, commonly at base, and normal grading, commonly in upper part, and (b) boundary structures, i.e. sole layer, sheared boundary, dewatering structures, lamination and stratification, sharp contacts, and inclusion of fragile megaclasts (Scott, 1988; Scott & *alii*, 1995), (4) sedimentary fabric of debris-flow deposits, clast roundness and composition, lahar bulking factors, and progressive downstream improvement in sorting, increase in sand and gravel, and decrease in clay. These downstream progressions are caused by incorporation (bulking) of better sorted gravel and sand (e.g., Osceola mudflow at Mount Rainier: Vallance & Scott, 1997).

Present-day debate bears on behavior and emplacement of debris flow, i.e., en masse emplacement as opposed to accretionary deposition (like progressive aggradation for emplacement of pyroclastic flows). Normal grading observed in the Osceola mudflow-deposits from Mount Rainier is best explained by incremental aggradation of a flow wave, coarser grained at its front than at its tail (Vallance & Scott, 1997).

The results and questions to be solved are used for the paleohydrological study of sediment gravity flows and to outline potentially hazardous areas to be affected by such flows, in particular by lahars: estimation of the cross-sectional area of flow, of the velocity, discharge, extent and volume of flows, and preparation of maps of inundation areas (e.g., Mount Rainier: Scott & *alii*, 1995; Scott & Sheridan, 1997). The distinction among cohesive lahar, noncohesive lahar, and debris avalanche is important for the purpose of hazard assessment, because cohesive lahars spread much more widely than noncohesive lahars that travel similar distances, and travel farther and spread more widely than debris avalanches of similar volume (Scott & Sheridan, 1997).

5. PROCESS-ORIENTED GEOMORPHOLOGY:
Denudation rates and geomorphic impact of substantial eruptions on volcanic landscapes

Volcanic geomorphology can contribute to the shift towards a process-oriented geomorphology through (a) development and use of accurate methods for measuring rates of geomorphic processes that shape ephemeral volcanic constructs, (b) identification of sources of material which contribute to sediment-delivery systems and measure the factors controlling the sedimentary budget on slopes and in valley channels draining active volcanoes, and (c) measurement and comparison of the geomorphic impact on selected catchments and their hydrologic response before, during, and after eruptions, which lead to the subsequent refinement of sedimentary and geomorphic parameters for the exponential decay model.

5.1. Denudation rates on volcanic landforms and on landscapes

A central question in volcanic geomorphology is how short- and long-term denudation rates of volcanoes have varied over time, in response to erosional episodes promoted by climate and sea-level changes, and tectonics.

5.1.1. On elementary volcanic landforms

Elementary volcanic landforms, like monogenetic cinder cones and single lava flows, have a clear starting time of geomorphic development, depending on climatic and physiographic factors. The evolution of tephra cones has long been the focus of morphometric studies (e.g., Kieffer, 1971, *in:* Cas & Wright, 1987). From statistical analysis on 38 cinder cones with a maximum age of 3 Ma from the San Francisco volcanic field in Arizona, Wood (1980) distinguished three different stages. With time and increasing diameter, cones show decreases in cone height, cone height/-cone basal diameter ratio and slope, but the ratio of crater diameter-cone basal diameter does not appear to change.

Incision rates in lava flow bedrocks measured in a variety of volcano-tectonic environments average 12.7 cm/ka (Righter, 1997). They range from as low as 0.5-8 cm/ka in Hawaï to as high as 23-25 cm/ka in the Atenguillo Valley, Jalisco (Mexico) and even up to 30 cm/ka in Utah. Comparable high incision rates are reported for areas undergoing extension. High incision rates may be characteristic of river channels at the edge of extensional regimes, because they are more likely to undergo base level changes due to normal faulting.

5.1.2. On composite landforms

Francis (1993) has summarized five stages in the erosional history of a volcanic stratocone. Karatson (1996) made use of the unique features of the Neogene/Quaternary volcanic chain in the Carpathians: age progression is reflected well in degraded stratovolcanoes from south to north in a similar moderate continental climate. Based on a complex morphometric analysis for 19 crater remnants dated from 11 Ma to 0.4 Ma, twenty-six variables have been examined by regression and factor analyses: numerical values of crater enlargment (109 m/Ma), internal valley growth (1.3 km/Ma), and average cone lowering (31.5 m/Ma) seem to answer the question of to what extent the stratovolcanoes are degraded. The calculated erosion rates seem suitable in worldwide comparison in moderate continental climate: the cone lowering rate 31.5 m/Ma fits well with global denudation rates inferred from other methods. Not surprisingly, time is the most important factor that explains the morphometric characteristics by about 40%. Adding two more factors, size-depth (crater-cone) and the distance from erosion base level, the stratovolcano morphology can be explained by 75 %.

5.1.3. Long-term denudation rate in volcanic landscapes

Short-term rates throw little light on the issue for determining how landscapes have evolved over enormous periods of time, which requires reliable chronological markers. In SE Australia, however, the widespread preservation of Tertiary basalts throughout the highlands and adjacent lowlands offers considerable scope for measuring the processes of wearing down and wearing back in the long-term denudation of a highland mass (Nott & *alii*, 1996). Both of these processes of denudation, dominance of scarp retreat versus summit lowering in the denudation of a highland mass, are found to be insignificant compared to the role of fluvial gorge extension over the last 30 m.y. Headward advancement of the Shoalhaven Gorge has been occurring at approximately 15 times the rate of major escarpment retreat (2500 m vs 170 m/Ma), 250 times the average rate of summit lowering, and 500 times the rate of interfluve consumption. Over the long term, the highlands in this region will become considerably more dissected well before they decrease substantially in height or are narrowed. The conclusion also has important bearing upon models predicting isostasic rebound from assumed character and rates of denudation.

5.2. Geomorphic impact of eruptions and sediment delivery on landscapes

Recent studies on volcanic landscapes affected by eruptions indicate that the geomorphic impact (erosion and sedimentation, and the subsequent recovery of disturbed watersheds) is complex, with an initial stage of accelerated erosion followed by an exponential decrease over a few years. However, the measured decreasing rate does not imply that erosion rates return to 'normal' pre-eruption rates.

5.2.1. Exponential decrease of the impact of eruption on landscapes

Most of the observations and studies of erosional processes are short-term and follow the first period after the eruption. Great volumes of clastic materials rapidly flood sedimentary environments and are supplied directly to basins, with sediment delivery and erosion rates increasing by two to four orders of magnitude above the pre-eruption rates for non-affected areas (Swanson & *alii*, 1983). However, erosion and sedimentation processes in volcanic landscapes affected by small to modest eruptions show a rapid decline, e.g., from 25-100 mm/yr in the first two years to 1-5 mm/yr within 5 years of the eruption of Mount St. Helens. Initial rates decrease with increase of infiltration in the ash layer, development of a vegetation cover and a steady drainage system.

5.2.2. Long-term effect of small to moderate eruptions

Following up the observations of Segerstrom (1950) on erosional and depositional processes that took place on the

Paricutin's cone and lava field, Inbar & *alii* (1994) emphasized the long-term effect and the recovery rates of the different landscapes 50 years after the eruption (1943-53). Three main periods of erosion were distinguished by Segerstrom: (1) accelerated erosion between the beginning of the eruption in 1943 and the end of the rain period of 1944; (2) a deceleration period between 1944 and 1952, and; (3) a gradient deceleration in erosion rates during the post eruption period until 1970, until they reach the normal values for the area. However, the erosion rates as of 1990 are about 50% above normal. The trend in the next period will be of slower rates of erosion and may extend over decades or centuries until a complete rehabilitation of the area.

In addition, rates of erosion and recovery will depend strongly on the different landforms in the area. On the cone and in the crater, erosion processes are slow as the vegetation cover grows rapidly. Most of the areas within the 25 cm isopach at about 8 km distance from the cone, are still covered with deep ash and unvegetated, thus exposed to erosional processes for a long term period. Floodplains are the most active area, where continous floods during the rainy period add sediments or induce the incision of channels along the edges of the lava fields. By contrast, lava fields are the least affected by erosional processes and most of the flows have a fresh appearance after 50 years. Thus, erosion processes and integration of drainage systems from basaltic lava flows are slow and may take centuries or thousands of years, as no integrated drainage is found in historically formed lava fields in different world climates (Inbar & *alii*, 1994, 1995).

5.2.3. Exceptionally delayed response to intracaldera eruption

Sediment response to tephra deposition from substantial volcanic eruptions, generally immediate and dramatic, can be delayed. Such important an exception is the 1886 eruption of Mount Tarawera, New Zealand, which blanketed > 200 km^2 with = 50 cm of scoria- and ash-fall deposits (White & *alii*, 1997). This fall deposit has suffered very little rilling or other surface erosion, and there was no immediate downstream redistribution of tephra by lahars or dilute floods. This lack of early response is attributed to the high permeability of the tephra deposit and gentle relief in the areas of substantial accumulation in caldera complexes of temperate to humid regions. Eighteen years after the eruption, collapse of a tephra bank that had controlled a raised intracaldera lake triggered a breakout flood, activating a decades-long period of intense tributary erosion in the upper catchment and damaging stream-bed aggradation outside the caldera. This abrupt increase in sediment yield from an initially stable posteruptive landscape has no documented precedents (White & *alii*, 1997).

5.2.4. Long-term sediment yield from impacted watersheds are poorly known

The impact of non-varying watershed parameters on erosion processes could be assessed over the Late-glacial and Holocene periods in the small Lac Chambon watershed (39 km^2, Massif Central, France), which exhibits forms inherited from the last glacier extension in ancient mountain relief without tectonics (Macaire & *alii*, 1997). Computation of stored material volumes and sediment yield values from plutonic and volcanic source rocks over the past 15,500 yr in the lake Chambon watershed gave accurate information: the mean mechanical erosion capacity by slope processes (16±6m) was 13 times greater than by running water (1.2±0.3 m), but it developed over only a quarter of the watershed surface. Fluctuations in sediment yield consisted of a 2.5-fold increase during cold and dry climate (Young Dryas) contrasting to a moderate decrease during humid climates (Pre-Boreal). A threefold increase in erosion over the last 1400 yr shows the impact of human-induced deforestation.

For the purpose of comparison, Macaire & *alii*, (1997) showed that erosion rates are higher in recent orogenies of mid- to high- elevation than in old orogenies. However, erosion rates as high as those computed over thousand of years can be attained over a few years only in volcanic watersheds affected by voluminous eruptions (= 10 km^3).

5.3. *Dramatic response from disturbed catchments to substantial eruptions*

Two cases are presented: the geomorphic effects of the 1980 eruption of Mount St. Helens and the first annual sediment budget (1980-81), and the extraordinary hydrologic response of the volcanic landscape following the 1991 eruption of Mount Pinatubo (The Philippines).

5.3.1. Assessment of the 1980-81 sediment budget around Mount St. Helens

Rosenfeld (1996) used a geomorphic approach instead of a hydrologic approach to assess the sediment budget around Mount St. Helens. First, potential sediment sources were identified, based on aerial photography and field investigations. Second, measurable landform units were categorized as well as numerous measurable temporary storage sites and sediment sinks. Third, a geomorphic classification of surface materials and landform types was constructed from post-eruption, repeated aerial photography. In addition, 65 control points were established and marked along the Toutle river valley, providing over 150 photogrammetric cross-sections of the constantly changing channel network.

Two types of sediment budget information were obtained (Rosenfeld, 1996). The total sediment storage was as high as 26 x 10^8 m^3. The net sediment yield was computed as 65 x 10^6 m^3 for the first year after the eruption: two thirds of which were delivered through erosion by the channel network at the expense of the debris-avalanche deposit. The geomorphic sources, volume, size, range and transport characteristics associated with storm events of different magnitudes were assessed. Besides lahars, most active erosional processes were shallow mass wasting, rill and bank erosion, and filling/breaching of detention ponds.

5.3.2. Protracted and worsening impact around Pinatubo

The response to large eruptions (= 10 km^3 of ejecta) had not been documented prior to the study of the Mount Pinatubo's lahars. The short-term impact around Pinatubo has been severe, but the mid-term impact appears protracted and worsening.

Impact and lahar occurrence

The climactic explosive eruption of Mount Pinatubo on June 15, 1991, which erupted a total bulk volume of 8.4 to 10.4 km^3, deposited abundant, loose 5 to 6 km^3 of pumiceous pyroclastic-flow deposits in the heads of valleys draining the volcano and about 0.2 km^3 of tephra on the volcano's flanks that would later be the primary source sediment for lahars. Numerous debris flows and hyperconcentrated flows were triggered during and following 1991 and affected 8 major drainages of Mount Pinatubo. They were triggered by (1) monsoonal rainstorms, sometimes enhanced by the passage of typhoons farther to the north, (2) volcanically induced convective rainstorms over localized heat sources, and (3) breakouts from debris dammed lakes. Although the areas affected by lahars have progressively expanded, the frequency of lahar events has decreased and the number of impacted river systems has dwindled to four in 1995, as source materials are gradually depleted. Lahars have been flowing into densely populated areas of central Luzon ove the past 6 years, taking a toll of lives, leaving more than 50,000 persons homeless, affecting more than 1,350,000 people in 39 towns and 4 large cities, and causing enormous property losses (= 1000 km^2 of prime agricultural land) and social disruption (Janda & *alii*, 1997).

Sources of material and flow types

Lahar sediment at Mount Pinatubo came from five distinct sources: (1) coarse tephra dropped from the eruption columns of the mid-June eruptions, (2) pyroclastic-flow deposits, (3) fine-grained tephra from ash-cloud deposits, phreatic explosions, and eruptive events postdating June 15, (4) 1991 lahar deposits, and (5) unconsolidated volcaniclastic deposits predating June 15 (Pierson & *alii*, 1997). Lahar deposition occurred primarily on low-gradient, coalescing alluvial fans 15 to 50 km downstream from the caldera at the base of the volcano, where deposit thicknesses generally ranged from 0.5 to 5 m. Total depositional volume on the east-side alluvial fans in 1991 was about 0.38 km^3, which is almost one-third of the potential contributing volume from the source pyroclastic sediments. Channelized lahars having peak discharges on the order of 100 to 1000 m^3/s typically were noncohesive pumiceous debris flows, some of which transformed to hyperconcentrated flows prior to final deposition. Flows range from turbulent, erosive hyperconcentrated flows to viscous, usually laminar debris flows (Phivolcs-Dost-Iavcei, 1995; Pierson & *alii*, 1997). Lahars in 1991 to 1994 were predominantly hot (50°C) and steaming, fed by sediments eroded from the thick pyroclastic-flow deposits filling valleys in the upper reaches of the watersheds. Succeeding lahars in 1995-96 were predominantly cold, and consisted of increasingly higher proportion of older, pre-1991 eruption deposits.

Sediment budget and yield

The sediment budget at Mount Pinatubo was evaluated by Pierson & *alii* (1992). In 1992, 2.3 km^3 of pyroclastic debris (about 40% of the 1991 pyroclastic flows) were bound to be delivered to rivers as long-term lahar deposits. A volume 3.9 km^3 of pyroclastic-flow deposits still remained in the watershed in 1992. At the end of the 1994 rainy season, about 2.2 km^3 has been already eroded from the 1991 pyroclastic-flow deposits and deposited on the alluvial aprons at the foot of the volcano (Phivolcs-Dost-Iavcei, 1995). Owing to the extraordinary thickness of the accumulated pyroclastic debris, occurrence of destructive lahars is expected to continue for several years. A first approximation of the potential yearly sediment budget, based on an exponential decay model was made by Pierson & *alii* (1992) and was one of the primary input towards constructing a lahar hazard map that can be utilized for long-term planning and rehabilitation of the Mount Pinatubo area. A decay rate intermediate between those of Mount St Helens and Mount Galunggung was chosen for Pinatubo. This exponential decay model is continuously refined as additional information became available through succeeding years, and lahar hazard maps are adjusted accordingly.

Sediment yields set world records during the first three posteruption years: sediment yields in 1991 were on the order of 1 million m^3 per km^2 per year, nearly an order of magnitude greater than the maximum sediment yield computed following the May 18, 1980, eruption of Mount St. Helens (Pierson & *alii*, 1997). In fact, the prodigious sediment yield from Pinatubo's upper and middle slopes and the sediment storage capacity in the adjoining lowlands are both diminishing, but at mismatched rates (Janda & *alii*, 1997). In general, sediment yields peaked early and are decreasing rapidly in east-side watersheds, where the volume of 1991 pyroclastic-flow deposits is relatively low, deposits and streams are confined in a few steep-walled valleys, thin ash fall from secondary explosions is common, and vegetation recovery is fast. Sediment yields peaked later and are decreasing slowly in west-side watersheds, where pyroclastic-flow deposits are more voluminous, numerous small streams drain a broad, gently-sloping, unconfined pyroclastic apron, and vegetation recovery was initially low.

The prodigious sediment yield from Pinatubo has also several aftermaths in terms of watershed disruption or piracy, channel avulsion, and blockage of tributaries. Blockage of tributaries at their confluence with the main channel, either by lahars or secondary pyroclastic flows, formed temporary lakes and impoundments. Floods and cold hyperconcentrated flows triggered by breaching of these naturally-dammed tributaries provided an additional hazard at Mount Pinatubo. Unlike rain-induced lahars, they can occur even in the absence of lahar-triggering rainfall, and therefore limit the capability to warn threatened areas (e.g., Pasig-Potrero in 1994: Phivolcs-Dost-Iavcei, 1995).

Cartographic modelling of erosion in the Sacobia's catchment

Prolonged intense rainfall associated with typhoons, and geomorphic accidents, such as secondary pyroclastic flow-induced stream piracy, and lake breakout events, can significantly alter the expected annual sediment delivery rate, as observed in the competing Sacobia-Abacan-Pasig Potrero river systems (Phivolcs-Dost-Iavcei, 1995). Detailed measurements and modelling should provide accurate information, relevant for improving both the exponential decay model and the lahar hazard maps. A cartographic modelling of erosion in pyroclastic-flow deposits of Mount Pinatubo (Daag & van Westen, 1996) encompasses the rapidly changing geomorphology of the Sacobia catchment on the eastern slope of the volcano before and during the eruption and for 3 consecutive years afterward. Emphasis was given to the importance of stream capture as a result of erosion and secondary explosions. To quantify the volumes of pyroclastic-flow material and annual erosion, five digital elevation models were prepared and analyzed using a GIS, and pre- and post-eruption geomorphological maps were elaborated. A total volume of 1.78 km^3 of pyroclastic flows deposited in 1991 in the Sacobia catchment covered an area of 24 km^2. Erosion rates were calculated to be in the range of 136-219 million m^3 per year, that is about 5.6 to 9.1 million $m^3/km^2/yr$.

The Pinatubo case study raises at least three important observations (Major & *alii*, 1997; Newhall & Punongbayon, 1997): (1) heavy rainfall alone was not responsible for generating the lahars in 1991. Another necessary condition was the alteration of watershed hydrology by volcanic deposits, prior, during, and after the eruption; (2) geomorphic 'accidents' affecting watershed and channels play a significant role in redistribution, reerosion, and redeposition of sediments. These will foster more lahars for the next 5 to 10 years, and more flooding beyond the alluvial fans onto the densely populated plains; (3) beyond the geomorphic impact of the devastating 1991 lahars, the subsequent lahars triggered by the seasonal monsoon rains and other geomorphic 'accidents' had far greater social and economic impact.

CONCLUSION: VOLCANIC HAZARDS, A RISING CHALLENGE

Increasing number of people at risk around hazardous volcanoes

At least 500 million people will be living under the shadow of a volcano by the year 2000 (Tilling & Lipman, 1993). Twice this century, large towns have been laid waste in minutes by volcanic eruptions (St Pierre, Martinique, in 1902 and Armero, Colombia, in 1985). Major population centers lie just tens of kilometers from several large volcanoes with a likelihood of eruption during the next century (e.g., Napoli near Vesuvius, Seattle and Tacoma near Mount Rainier, Manila near Taal). An indication of volcanic risk in selected countries can be gleaned by comparing projected urban population densities at the end of this decade with the areal distribution of volcanoes (Pyle, 1995).

The problem is of wider significance, because «good science alone will not do the job of reducing volcano risk» (Tilling & Lipman, 1993). Although there have been many advances in our understanding of factors leading to the occurrence of eruptions during the past decades, the ability to predict volcanic events is still poor. Research needs range from fundamental investigation of the causative processes to direct scientific and engineering studies to mitigate risk. Increasingly, policy decisions are being based on a formalized «risk assessment» that requires a scientifically based probabilistic determination of the likelihood of a natural hazard event occurring. Such probabilistic studies are only as good as the data and knowledge input into them.

How can volcanic geomorphology contribute to hazard and risk assessment?

Geomorphological surveys form a logical starting point for natural hazard zoning (Verstappen, 1988). Process-oriented geomorphology foster primary input for quantitative reconstruction of recent volcanic activity, and for the development of models used in long-term planning (Rosi, 1996). Additional risk assessment and zonation requires the development of a series of scenarios in which eruption magnitudes, hazard types, composite risk zonation indices, and the vulnerability of people and infrastructure are adequately considered (Blong, 1996).

Geomorphology can contribute to risk assessment through two approaches: geomorphic hazard zonation and composite risk zonation (Slaymaker, 1996). The procedure improves previous geomorphological methods and leads to a risk zonation encompassing three classes at Mount Pinatubo (Nossin & Javelosa, 1996). Geomorphic hazards, both volcanic and non-volcanic, were identified and analysed with the aid of satellite imagery and field survey. Geomorphic hazard domaines were established according to the capacity of each hazard to affect geomorphic stability, the perceived people vulnerability, and the priority of georesource function. A composite risk zonation, incorporating geomorphic mapping, geomorphic risk analysis and georesource priority, was calculated.

However, the assessment of, and the response to, unprecedented hazardous situations remain a continuing dilemma for geoscientists, engineers, residents, and policymakers (Janda & *alii*, 1997).

Abstract: Thouret J.-C., *Research problems and potential contributions to related earth sciences: an overview.*

The session «Volcanic Geomorphology» of the Fourth International Conference on Geomorphology (Bologna, Italy, 28 August-3 September 1997) has examined what was the role of geomorphology in analyzing the volcanoes on Earth. Volcanism has a considerable geomorphological significance, because it directly creates and degrades landforms, and provides an age either for the land surface over which the erupted material lies or for the succession in which it is intercalated.

Volcanic geomorphology enables us to (1) reconstruct original forms even after significant erosion has occurred, (2) estimate denudation rate over a known period of time, based on geochronology and reconstruction techniques, and (3) compare the effects of climate on the nature and rate of denudation, because similar volcanic landforms and rocks are found over a wide range of climates.

Volcanic geomorphology aims to (1) decipher the complex interplay of eruptive and geomorphic processes, and the distribution and source of volcaniclastic sedimentation, (2) unravel the influence of volcanism, tectonics, and paleoenvironments on landscape, and (3) assess short- and long-term response to sediment delivery from substantial eruptions, and subsequent geomorphic impact in disturbed watersheds.

Finally, volcanic geomorphology should face the rising challenge posed by the combination of natural hazards and *increasing number of people at risk around volcanoes*. Volcanic geomorphology can contribute to risk assessment through geomorphic hazard zonation and composite risk zonation.

Key Words: Volcanoes, Landforms.

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COMMENTS ON POSTERS AND ORAL PAPERS PRESENTED

The topics dealt with in this scientific session of the IV International Conference on Geomorphology are of very various kind and clearly express the recently increasing interest towards geomorphological studies in volcanic areas. Really, some of the papers and posters presented are not strictly about Volcanic Geomorphology and probably would have found a more suitable place in other sessions. But the vast amount of contributions presented at an International Conference and the sometime eccessive conciseness of the abstracts submitted not always allow an optimal sharing into the different sessions.

Actually, when examining thoroughly the paper contents a question may arise spontaneously: what does really mean «Volcanic Geomorphology»? or rather «which are the boundaries of this branch of Geomorphology»?

To give a full answer to this question it would be necessary a long discussion among researchers involved in Volcanic Geomorphology. It would be presumptuous only thinking it possible to suggest an answer in few words; on the other hand it is far beyond the convenor's task which is simply to evidence the specific topics the posters deal with, in order to make easier their consultation and to favour their discussion.

A first answer about the *main points and importance of Volcanic Geomorphology studies* is offerred by the poster of general content by M. Inbar on the «New trends in Volcanic Geomorphology». In fact the Author underlines textually that: «a main theme in Geomorphology is the evolution of landscape and volcanic areas offer the opportunity to monitor landscape evolution from its beginning» and again «volcanic landscapes are characterized by two common features: there is a clear starting time of the geomorphic development and the lithology is similar in the different climatic areas of the world».

In this perspective, the significance of geomorphological studies on volcanic areas is evident. By comparing the preservation degree of volcanic landforms from different parts of the world, it is possible to estabilish their relative dating and to evaluate the influence of different structural and climatic conditions on erosion rate and relief evolution. Moreover, considering the rate of relief building due to active volcanism, the way the constructive processes interfere with the destructive exogenous ones can be recognized with less difficulties.

In spite of the large variety of the topics examined by the other posters, some general themes can be tentatively identified.

A first theme concerns the relations between *Morphology and tectonics in volcanic areas.* Volcanic areas are an excellent place where the interplays between morphology and tectonics can be successfully examined. In fact morphological indicators for tectonics can be found not only among the landscape features originated by exogenous agents, but also among volcanic landforms, the location and shape of which are unmistakably tied to regional and local tectonic dislocations. On this basis morphotectonic investigations can aid to single out the tectonic conditions which favour or have favoured the volcanism development, to reconstruct the eruptive sequence and to delineate the morphological evolution of the volcanic reliefs.

Three posters face such topic and all of them consider Italian volcanoes. Two of them, (Ciccacci & *alii*), are about inactive volcanoes of Central and Southern Italy respectively and stress the role of geomorphological studies in the reconstruction of the volcanic history of inactive volcanoes (Roccamonfina and Monte Vulture); moreover they evidence the usefulness of the quantitative approach in the definition of the tectonic setting in which volcanism developed The third poster «Volcanisme et tectonique sur la façade orientale de l'Etna», by Kieffer, underlines the marked control of regional tectonics on the eruptive history of this volcano and on its morphological evolution, which is tied also to impressing phenomena of gravitative tectonics following a mechanism which reseambles the one responsible for some particular structures of the Hawaiian volcanoes.

The second theme regards the important problem of the *Erosion processes in volcanic areas.* The interest of this topic is twofold: one is purely scientific and the other is of applicative kind. In the areas where the effects of volcanism are manifest and the role of tectonics on the relief shaping is more easily discernable, as a consequence the denudational process dynamics can be better understood. On the other hand erosion process on active volcanoes are often hazardous processes and therefore the knowledge of their dynamics is very important in the risk reduction.

The group of posters dealing with these important problems is the largest one. The first two are complementary and consider the occurrence of debris flows on the slopes of active volcanoes. The poster «Sediment discharge by storm runoff at volcanic torrents affected by eruption» (Suwa & Yamakoshi) underlines that the decrease in infiltration capacity due to the newly emplaced tephra and pyroclastic deposits is a main cause for high erosion and debris flow triggering; and this is proved by data relevant to three different Japanese and Indonesian volcanoes which show that permeability increases and erosion rates decrease with increasing time after the eruption. The second poster, (Onda & *alii*), underlines that the permeability decrease soon after the eruption of Unzen Volcano is tied to physico-chemical processes of the ash which led to the formation of a chemical crust; moreover it indicates that use of electrolyte can increase the permeability and reduce the surface runoff and therefore the debris flow frequency.

The poster by M. Hürliman & *alii* considers landslide events which are significant processes on volcanic edifices. Through a Geographycal Information System, the links between initiation of landslide and calderic collapses in Tenerife are investigated. The numerical model obtained in this way undeline that fracturing, horizontal tensile stress and seismicity can surpass the stabilising forces of a volcano flank and trigger large landslides.

The erosion by channelled waters and the relevant modifications on the volcanic landforms is the object of the poster, (D. Karatson), on erosional craters and caldera, the genesis of which is here tied also to climatic conditions; in particular a pluviometric threshold is evidenced for the development of amphitheatre valleys which should occur only when annual rainfalls exceed 1,500 mm.

Landforms due to effusive magmatic events and exogenous processes on the northeastern slope of M. Etna is the object of the following poster, in which Congiu & *alii* evidence, among other things, the possible existence of glaciers on M. Etna during the last glacial maximum. Beyond the geomorphological interest this possibility might serve as a starting point to better understand the Quaternary climate evolution of the Mediterranean area.

The object of the last poster of this group is the morphology of a seamount offshore the coast of Campania (Gabbianelli & Colantoni); the morphological features surveyed allowed, besides other things, to recognize that the morphological evolution has been strongly conditioned by Late Pleistocene glacio-eustatic variations which determined the emersion of the seamount summit and triggered widespread erosion processes .

Another important theme concerns directly the *Hazards in volcanic areas*, a topic which has already been touched on by the previous papers about debris-flow occurrence and reduction. Geomorphological studies in this applicative field are a useful tool to single out areas prone to the most hazardous events, as they help the estabilish the type and sequence of the volcanic activities.

In the first poster of this group the Late Pleistocene eruptive history of Misti Volcano, in the Central Andean Volcanic Zone, is reconstructed in details by means of fieldwork, air-photos and Spot satellite images (J.C. Thouret & *alii*). On the basis of the eventful eruptive history of this mainly explosive volcano and of the deposit distribution, the most severe and probable hazards are indicated.

The second poster is of different kind because it does not deal strictly with volcanic hazards but with the possible hazard due to soil erosion in a typical volcanic area like Iceland surely is (Käyhkö & *alii*). The authors underline that soil accelerated erosion started after human settlement and is associated to intense eolian processes; moreover they recognize that lava morphology can significantly control the eolian transport and deposits, with

The two posters left do not face a specific theme, but underline the importance of the volcanic product outcrops in the geomorphological reconstructions. The first poster (D. Chandrasekharam & H.C. Sheth) on the erosional history of Deccan flood basalt shows that studies about the stratigraphy of volcanic products can aid the reconstruction of the geomorphological evolution of the study areas starting from the pre-volcanic landscape. In the second poster, petro-structural and geochemical analyses of volcanic products permit to single out an ancient and morphologically undetectable volcanic zone (M. Barbieri & *alii*).

This synthesis, although concise, evidences the vast contribution of knowledge on Volcanic Geomorphology and the numerous methodological hints that the papers presented as poster contain, thus confirming the existence of promising bases for the development of this branch of Geomorphology.

This view is strengthened also by the papers object of oral presentation which are worth mentioning for their interesting contents. The first paper, by M.N. Zimmerman, expresses the extent to which volcanic activity can modify the morphology of wide areas. It shows the morphological changes in the basin headwaters and in the river lower reaches which followed the 1991 eruption of Mount Pinatubo, Philippines. In particular, the pyroclastic flows erupted and the ones produced by secondary phreatic explosions, due to the infiltration of rainwaters, filled whole valleys and caused deep changes of the drainage networks in the headwaters. However, lahars tied to the occurrence of typhoons caused the most dramatic modifications; huge volumes of sediments were transported in the lower river reaches and wide agricultural lands were covered by lahar sediments, which implied enormous socio-economic impacts.

The paper by M.J. Rossi examines the morphological aspect of postglacial lava flow fields in Iceland. Three main types of lava flow fields are recognized which differ for the morphology of their surfaces: monogenetic shield volcanoes characterized by shelly-type pahoehoe, regional lava flows from volcanic fissures having an aa surface texture and lava flows from central volcanoes completely lacking in pahoehoe structures. The causes of the observed morphological differences appear to be tied not only to the varying rheological properties of the lava but also to the supply rate at the vent and the amount of branching of the lava flows.

The paper presentation and the poster session has been followed by a discussion focussed mainly on the oral communications, as the length of the poster session allowed the direct and fruitful dialogue between authors and interested people.

Finally it is worth mentioning that at the end of the discussion the possible constitution of a study group about Volcanic Geomorphology has been proposed by Nossin; successively the proposal has been approved by the International Association of Geomorphologists.