Erosion calderas: Origins, processes, structural and climatic control

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Abstract The origin and development of erosion-modified, erosion-transformed, and erosion-induced depressions in volcanic terrains are reviewed and systematized. A proposed classification, addressing terminology issues, considers structural, geomorphic, and climatic factors that contribute to the topographic modification of summit or flank depressions on volcanoes. Breaching of a closed crater or caldera generated by volcanic or non-volcanic processes results in an outlet valley. Under climates with up to ~2000–2500 mm annual rainfall, craters, and calderas are commonly drained by a single outlet. The outlet valley can maintain its dominant downcutting position because it quickly enlarges its drainage basin by capturing the area of the primary depression. Multi-drained volcanic depressions can form if special factors, e.g., high-rate geological processes, such as faulting or glaciation, suppress fluvial erosion. Normal (fluvial) erosion-modified volcanic depressions the circular rim of which is derived from the original rim are termed erosion craters or erosion calderas, depending on the pre-existing depression. The resulting landform should be classed as an erosion-induced volcanic depression if the degradation of a cluster of craters produces a single-drained, irregular-shaped basin, or if flank erosion results in a quasi-closed depression. Under humid climates, craters and calderas degrade at a faster rate. Mostly at subtropical and tropical ocean-island and island-arc volcanoes, their erosion results in so-called amphitheater valleys that develop under heavy rainfall (> ~2500 mm/year), rainstorms, and high-elevation differences. Structural and lithological control, and groundwater in ocean islands, may in turn preform and guide development of high-energy valleys through rockfalls, landsliding, mudflows, and mass wasting. Given the intense erosion, amphitheater valleys are able to breach a primary depression from several directions and degrade the summit region at a high rate. Occasionally, amphitheater valleys may create summit depressions without a pre-existing crater or caldera. The resulting, negative landforms, which may drain in several directions and the primary origin of which is commonly unrecognizable, should be included in erosion-transformed volcanic depressions.

Key words Erosion caldera · Erosion crater · Degradation and drainage of volcanoes · Volcanic geomorphology

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Introduction

Erosionally formed volcanic depressions, which may be several kilometers in diameter, are traditionally referred to as erosion calderas. The origin and development of such depressions have not been studied in detail by volcanologists. Even geomorphologists have paid little attention to the formation and evolution of erosion calderas. In general, volcanic landscapes and their evolution are studied infrequently (cf. Francis 1993). As a consequence, several terms for erosional depressions in volcanic terrains are used in the literature (Thouret 1993, in press).

A traditional explanation of erosion calderas, widely accepted presently, has been given by Ollier (1988). His description and examples of erosion calderas are almost the same as those in the most comprehensive book on volcanic landforms by Cotton (1952). This explanation envisions a pre-existing, closed crater or caldera that is later breached and eventually enlarged by a radial valley that develops on the external slopes of the volcano. Such an interpretation implies that fluvial erosion is the major process determining the subsequent shape of the volcanic depression. A frequently quoted example of such a basin is Caldera La Palma, presently named Caldera de Taburiente, Canary Islands (Ollier 1988). This depression, from which Charles Lyell took the word *caldera* for general application, was thought for more than a century to have been enlarged primarily by headward erosion (Lyell 1855; Middlemost 1970). Recent studies, however, attribute the origin of Caldera de Taburiente mainly to giant landslides (Staudigel and Schmincke 1984; Ancochea et al. 1994; Masson 1996), so that it is no longer the exemplary type of caldera formed by fluvial erosion.

Much has been learned recently about the significance of volcanic slides and debris avalanches that frequently occur on active and dormant volcanoes all over the world. As a result, we now can distinguish between closed or quasi-closed breached depressions called erosion calderas and the open or half-open avalanche, amphitheater, or horseshoe calderas (Siebert 1984). Erosion calderas are formed by long-term, small-scale fluvial processes; horseshoe-shaped calderas are formed by short-term, large-scale debris avalanches or may be created or transformed under cold and wet climate by glaciers, such as at Mount McLoughlin, Oregon (Smith 1990).

Despite this restriction, the origins of many erosion calderas may be uncertain or, when known, polygenetic. For example, a wide range of volcanic, tectonic, and erosional processes may modify the subsequent shape of a volcanic depression.

In this paper we review the common ways in which erosion calderas form and evolve, the factors that control their creation, and how can they be classified.

Degradation and drainage of volcanic edifices

The flanks and summit region of volcanic landforms that are positive and concentric are expected to be deeply dissected and degraded by fluvial erosion. Constructional edifices, such as composite and shield volcanoes, pyroclastic cones, and lava domes, are typical. These forms must be built, at least in part, of impermeable rocks to enable the drainage network to develop. Permeability depends on many factors, mostly endogenic ones (geological structure, surface texture, rock type, welding, etc.), which are accentuated by exogenic agents such as rainfall and frost action. Even elementary volcanic landforms show great variability. For example, fresh scoria cones, built of permeable material, are usually not liable to fluvial erosion (Bloom 1991), whereas spatter cones, or those composed of welded scoria fall or flow deposits, experience intense gullying. However, gullying can occur even on coarser-grained material, such as scoria, when rainfall rate is greater than infiltration, because erosion depends also on rainfall intensity (Collins and Dunne 1986).

After cessation of volcanism, a cone or dome is exposed to rapid fluvial erosion. The typical course of degradation (Cotton 1952; Macdonald 1972; Ollier 1988; Francis 1993) is reviewed herein. Significant erosion can also take place during the eruptive period (Moriya 1987; Cas and Wright 1988; Carracedo 1994):

Fig. 1 Parasol ribbing on the external slopes of the active Sakurajima volcano (Kyushu, Japan)

Fig. 2 Minor fissure formed during the 1959 phreatic activity of Mt. Shinmoedake, Kirishima volcano (Kyushu, Japan), has initiated gullying in the neighboring slopes. (Courtesy R. Imura, Kagoshima University)

1. In the first, "initial" stage, rill erosion of gullies or barrancos dissects the surface and creates a radial (centrifugal) drainage pattern on the external volcano slopes by parasol ribbing (Fig. 1; Cotton 1952). Ash avalanches (e.g., on Vesuvius; Cotton 1952; Hazlett et al. 1991) can accelerate this process early, depending on material, climatic, and topographic conditions, or wave erosion such as at Kaho'olawe (Hawaiian Islands; García 1990) and Volcán Ecuador (Galápagos Islands; Rowland et al. 1994). Volcanic initiation of slope transformation can be observed on Mount Shinmoedake, Kirishima volcano (Japan). A minor fissure and the deposition of fine-grained ash, generated during phreatic activity in 1959 (Imura and Kobayashi 1991), have initiated gullying in the surrounding area (Fig. 2). On Lascar volcano, Chile, the 1993 pyroclastic flows produced intense erosion (Sparks et al. 1997), showing that initial incision of volcano slopes can be caused primarily by eruptions.
2. In the second, "mature" stage, major water courses that collect the initial gullies start to form (Fig. 3).

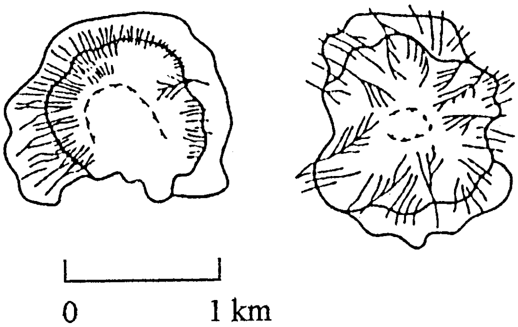


Fig. 3 Formation and reduction of drainage on simple scoria cones in the Cima field, California. *Left:* cone BB (0.33 Ma); *right:* cone K (0.99 Ma). Outer belts are the aprons; *dashed lines* indicate crater rim. (Modified from Dohrenwend et al. 1986)

This process, called channel capture, is controlled in the short term by rapid valley incision and beheading and by increased permeability of the eroding material (Horton 1945; Dunne 1980; Collins and Dunne 1986; Dohrenwend et al. 1986). Channel capture is well documented either under semiarid climate (Cima volcanic field in California; Dohrenwend et al. 1986) or under temperate to tropical climate (the Michoacán-Guanajuato area in central Mexico; Hasenaka and Carmichael 1985). In the Cima field, where vegetation recovery is slow, channel excavation is accelerated by debris flows down the initial gullies, and piping may also contribute to intense gully development (Dohrenwend et al. 1986). In the long term, the reduction of the drainage network is related to vegetation recovery, which hinders fluvial erosion. Under climates having sufficient precipitation (ca. 600–1500 mm annual rainfall), dense vegetation allows a smaller but constant number of channels to evolve (Ohmori 1983). With higher precipitation, erosive force surpasses plant protection against erosion. Effects of heavier rainfall on volcano degradation are discussed later. Under all climates, and mostly on larger cones, channel capture may result occasionally in the formation of triangle-shaped remnant surfaces of the edifice between the incising valleys. These triangle surfaces were termed "planèzes" by de Martonne (1950).
3. In the third, "residual" stage, primary landforms and original surfaces disappear; the volcanic origin is indicated by the radial drainage network only. In the center of the volcano, high-rate degradation of cinder or composite cones lasting 1–3 Ma may expose the core complexes (vent structures) (Francis 1983; Dohrenwend et al. 1986).
4. Finally, in the "skeleton" stage, all volcanic products and landforms are degraded. Only rocks of the subvolcanic level, cropping out as necks, dykes, and

sills, may be preserved for a longer time and rise above the surroundings (Mt. Taylor volcanic field, New Mexico; Tibesti Mountains, Chad; Skye and Mull in British Tertiary Volcanic Province, Scotland; Cantal, Auvergne, France).

Volcanoes in different climates do not follow the described erosion scenario in the same way and at the same rate. Kear (1957) found that, in New Zealand, the planèze stage can be traced back to Middle Pleistocene, residual stage to Plio-Pleistocene, and skeleton stage to Upper Miocene. King (1949) described a Lower Miocene volcano from Uganda with preserved planèzes; similar landforms can be traced back to 15 Ma in the Carpathians (Karátson 1995) and to 6–9 Ma in the Cantal, Auvergne, France (de Goër de Herve et al. 1985). Erosion is much more intense under heavy or concentrated rainfall. For example, 0.4- to 0.7-Ma Japanese volcanoes, such as Hakone and Oguni, Central Japan, have already lost their craters and planèzes (Suzuki 1969). In tropical areas of Indonesia, erosion needs only 20,000 years to reach the residual stage (Ollier 1966). Apart from wet climate, high-rate exogenic processes, such as glacial erosion or periglacial frost action, also accelerate degradation (Aleutian Islands, Alaska; High Cascades, Canada–United States; Bárdarbunga, Grímsvötn, Iceland; Beerenberg, Jan Mayen Island; Ardèche and Auvergne, France; Mt. Kenya, Kenya; Kilimanjaro, Tanzania; Ross volcano, Auckland Islands).

The amount and intensity of rainfall influence to a great extent whether planèzes can form at all. Planèzes are described from the Hawai'i, Lesser Antilles, Central Africa, La Réunion, St. Helena, and Auvergne volcanic areas, locations characterized by heavy or significant rainfall in tropical, subtropical, or oceanic climate. In contrast, planèzes are scarce or hardly preserved in semiarid to temperate continental climates. With such low rainfall, the drainage network develops slowly enough to degrade the whole cone surface uniformly; hence, volcanoes may reach the residual stage without leaving sizeable original surfaces between valleys. Slow erosion rate then may enable craters to be preserved even until the residual stage, as for Miocene volcanoes in the Carpathians (Karátson 1994, 1996).

On the other hand, several factors other than climate, such as geological setting, rock composition, eruption sequence, volcano size, volcano stability and structure, rock alteration, relative and absolute edifice height (determining a possible ice cap or snowpack), distance from erosion base level, and wind direction, may be responsible for different degradation styles which may occur on the same volcano (Bullard 1962; Ollier 1988; Mahaney 1989; Mills 1993; Karátson 1996). The edifices may, at least partly, be degraded before the core complex is exposed, or the core may come to the surface in an early stage of degradation. We emphasize that intense initial erosion, up to several thousand millimeters per thousand years (Ruxton and McDougall 1967; Ollier and Brown 1971; Ollier 1988; Mairine and Bachèlery 1997), may slow dramatically once vegetation covers the entire surface (Segerstrom 1950; Inbar et al. 1995). Under temperate continental climate in the Carpathians, volcanoes may be eroded at a rate of no more than 30 m/Ma for tens of millions of years (Karátson 1996).

A database of selected erosion calderas presenting volcano type, age, elevation, depression diameter, as well as annual temperature and rainfall values is given in Table 1. Table 1a includes various examples of erosion depressions at extinct volcanoes where erosion is an evident factor in modifying the primary shape. Table 1b includes extinct or dormant, breached, primary volcanic depressions the shapes of which have been only slightly modified by erosion. Obviously, the two groups cannot be distinguished clearly.

All the examples, except for Veniaminof caldera (Alaska), are from volcanic terrains in a fluvial regime. The exception is very important: Veniaminof has undergone rapid modification by glacial erosion. It is a medium-sized, ice-filled caldera with a broad, glacier-cut primary breach, a probable secondary subglacial breach, and numerous glacier-cut notches in the rim (Yount 1990). Veniaminof argues again for the restriction of the term erosion caldera to fluvially breached and eroded depressions, the rims of which remain unindented and circular and the "caldera" shape of which remains recognizable for a long time.

Table 1 contains two groups of depressions, those with one outlet valley and those with more than one (Karátson 1997). The first group includes generally smaller but more regular-shaped, unidirectionally breached depressions the origins of which can be easily inferred to be a summit crater, summit caldera, caldera volcano, etc. The second group contains generally larger and more irregular depressions, somewhat similar to glacier-modified depressions. The characteristics of these two groups are given herein, following a summary chart (Fig. 4) that also displays sector-collapse horseshoe calderas and glacier-modified calderas.

Depressions with one outlet valley

We distinguish between five types of unidirectionally breached erosional depressions in volcanic terrains, focusing on what happens in and around the center of the volcano.

Erosion of simple craters of cones or domes

During the first stage of erosion, a dense, centripetal drainage develops on the free surfaces inside the crater. This gully network may decline or even disappear after a relatively short time interval, approximately tens of thousands of years. The decline of drainage density inside the crater, which is a function of both time and climate, is presented in Fig. 5. We can see intense gullying

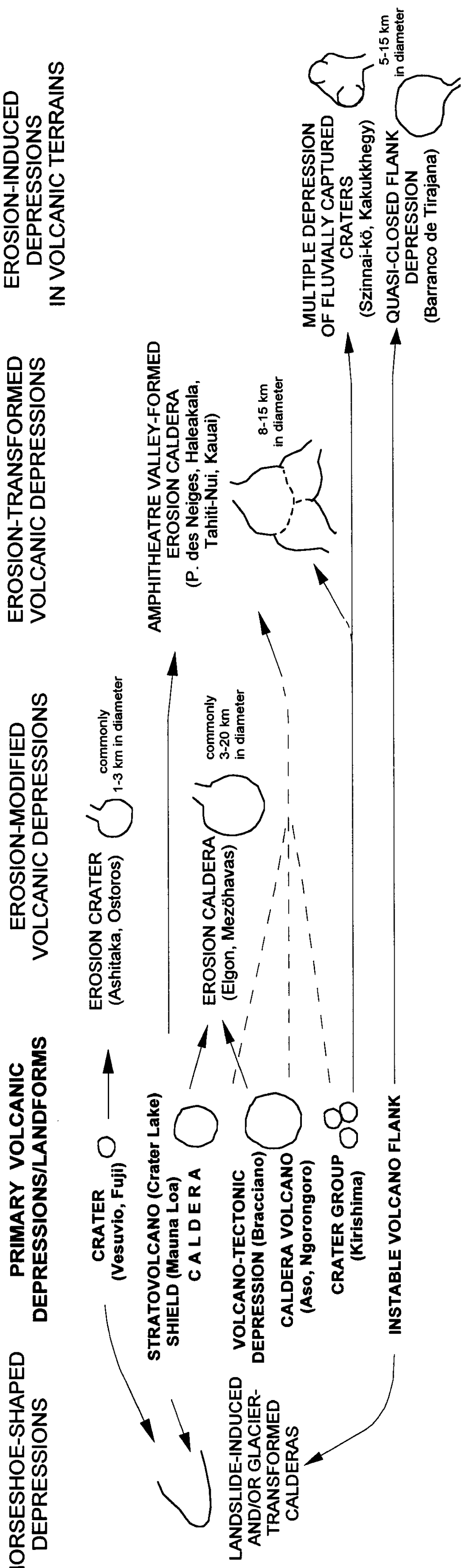


Fig. 4 A classification of breached depressions in volcanic terrains

in the crater of the active cone of Vulcan, Papua New Guinea (annual rainfall 2200 mm). In contrast, the 1500-m-wide western crater of the Ciomadul (Csomád) volcano, East Carpathians (Romania), which last erupted 30–40 ka (Moriya et al. 1996), is characterized by a ca. 1000 mm annual rainfall and is dissected very little and has no breach.

In a more mature stage of erosion, unified major water courses retreat by headward erosion on the outer slopes of the cone or dome. Initial fractures and topographic lows in the rim – caused or enhanced by volcanic activity (lava flows, pyroclastic flows), geothermal activity, different rate of alteration, small-sized landslides, debris flows, etc. – may influence where the breaching occurs. Given no primary breach and all other factors being equal, one of the external channels that has the highest valley-forming energy is eventually able to breach the crater rim. The erosive energy of a valley is determined mostly by the slope angle, the erosion base level, and the drainage basin area. If the newly breached crater contains a lake, small-scale landslides and related overflows can accelerate draining of the lake (such as Maughan lake, Mt. Parker volcano, southern Mindañao; Global Volcanism Network Bulletin 1995). For a peat-bog (sphagnum) fill, the convex surface of the peat-bog may localize the first tributaries of the breaching channel between the margin of the bog and the crater rim (such as at Mohos crater of Ciomadul volcano, East Carpathians, Romania).

Invasion of one of the external channels into a closed crater is a turning point in the development of the drainage network and the erosional transformation of the crater. The drainage basin of the breaching channel rapidly increases by adding the area of the crater. The resulting channel gains much energy relative to that of the other valleys on the outer slopes. The initial centripetal drainage becomes dendritic (Ollier 1988), clearly showing that the internal tributaries now feed the outlet valley. This process, referred to as *unidirectional breaching*, is related to normal erosion valleys that typically form under temperate climates. Dunne (1980) recognized that catchment areas increase mainly by headward erosion until a threshold is reached in drainage network development. In a mature stage, however, channel lengths do not change significantly; hence the drainage pattern remains "fixed." Oguchi (1995) pointed out that, starting from an initial pattern, this stage may have been reached in the mountains of Central Japan, with an annual rainfall of 1500–2000 mm, since the last glaciation. Young craters (<1 Ma) of that area (e.g., Bishamon, Dainichi, Azumaya; Shimizu et al. 1988) can have a diameter as large as those of the oldest (10–11 Ma) Carpathian craters (Karátson 1996), but they also have only one outlet valley. Moreover, one can see unidirectional breaching in a region of even higher precipitation, such as the Phil-

Table 1a Topographic, morphometric, and climatic data for selected erosion depressions in volcanic terrains around the world, with regions in ascending order of precipitation. *UB* unidirectionally breached; *MB* multidirectionally breached. Note the elevation differences between the highest points and the meteorological stations. Each additional 100-m increase in elevation corresponds to ca. –0.6 °C annual temperature drop and variably increasing (sometimes decreasing) rainfall, also controlled by exposure and orography

Volcano (highest point, m)	Country	Geographic position	Known or most likely origin of the present depression	Cessation (youngest age) of volcanism	Depression diameter (m)	Name and geographic position of the nearest meteorological station	Station elevation (m)	Annual temperature (°C)	Annual rainfall (mm)	Reference
The Carpathians										
Ostoroş (Osztorόc, 1385)	Romania	46°34′N/25°38′E	UB stratovolcanic crater	6.3 Ma	2360	Sfintu Gheorghe 45°52′N/25°47′E	561	7.6	584[a]	1–3
Rotunda (Kerek-hegy, 1241)	Romania	47°86′N/23°65′E	UB stratovolcanic crater	10.3 Ma	2790	Sfintu Gheorghe 45°52′N/25°47′E	561	7.6	584[a]	2
Seaca-Tătarca (Mezőhavas, 1777)	Romania	46°41′N/25°13 E	UB stratovolcanic or shield caldera	7.0 Ma	5250	Sfintu Gheorghe 45°52′N/25°47′E	561	7.6	584[a]	1, 3
Strechov (Sztrehó, 915)	Slovakia	48°47′N/21°33′E	UB stratovolcanic crater	10.7 Ma	2720	Presov 49°00′N/21°15′E	270	8.3	631[a]	2, 4
Pol′ana (Polyána, 1458)	Slovakia	48°38′N/19°29′E	UB stratovolcanic caldera	12.5 Ma	6400	Presov 49°00′N/21°15′E	270	8.3	631[a]	4
North America										
Mt. Taylor (3471)	USA (New Mexico)	35°15′N/107°35′W	UB stratovolcanic caldera	1.5 Ma	~4500	Albuquerque 35°03′N/106°37′W	1620	13.8	739	5
The Mediterranean										
Sacrofano (434)	Italy	42°′N/12°21′E	UB caldera volcano	0.3 Ma	~6000	Roma 41°54′N/12°29′E	46	15.6	881	6
East Africa										
Mt. Elgon (4321)	Kenya/Uganda	1°10′N/34°35′E	UB stratovolcanic caldera	?Pliocene	~8000	Kisumu 0°07′N/34°35′E	1158	23.1	1111	7
Japan Islands										
Kariba (1520)	Japan (Hokkaido)	42°37′N/139°55′E	UB stratovolcanic and/or landslide caldera	~0.5 Ma	4030	Hakodate 41°49′N/140°45′E	35	8.5	1170	8
Ashitaka (1457)	Japan (Honshu)	35°13′N/138°48′E	UB stratovolcanic crater	700 ka	1750	Tokyo 35°41′N/136°46′E	6	14.3	1575	8
Dainichi (1709)	Japan (Honshu)	36°00′N/136°47′E	UB stratovolcanic crater	0.94 Ma	2380	Tokyo 35°41′N/136°46′E	6	14.3	1575	8, 9
						Wajima 37°23′N/136°54′E	6	12.7	2278	
Azumaya (2333)	Japan (Honshu)	36°35′N/138°24′E	UB stratovolcanic crater or caldera	0.60 Ma	3340	Tokyo 35°41′N/136°46′E	6	14.3	1575	8, 9
						Wajima 37°23′N/136°54′E	6	12.7	2278	
The Philippines										
Mt. Sibulan (1322)	The Philippines (Mindañao)	6°37′N/125°25′E	UB stratovolcanic crater or caldera	?Late Pleistocene	3200	Davao 7°01′N/125°35′E	23	27.0	1928	–
						General Santos 6°07′N/125°12′E	<50	26.9	943	
Mt. Isarog (1976)	The Philippines (Luzon)	13°40′N/123°23′E	UB stratovolcanic crater	Holocene	2250	Legaspi 13°09′N/123°45′E	6	27.0	3371	10
						Lucena 13°55′N/121°37′E	<50	27.1	1931	
Misamis Oriental (2440)	The Philippines (Mindañao)	8°45′N/125°48′E (highest point)	UB, probably multiple stratovolcanic caldera with amphitheater valleys on the W, N, and NE slopes	?Late Pleistocene	~9000	Malaybalay 8°21′N/125°07′E	642	23.6	2736	–
						Dipolog 8°36′N/123°21′E	<50	27.4	2921	
Misamis Occidental (2404)	The Philippines (Mindañao)	8°18′N/123°38′E (highest point)	Large amphitheater valleys with no recognizable crater/caldera	?Late Pleistocene	–	Malaybalay 8°21′N/125°07′E	642	23.6	2736	–
						Dipolog 8°36′N/123°21′E	<50	27.4	2921	
Pacific Ocean										
Haleakala (3055)	USA (Maui, Hawaii)	20°43′N/156°15′W	MB shield caldera	0.4 Ma[b]	~7900	Haleakala 20°46′N/156°15′W	2142	11.8	1350	11
						Kipahulu 20°39′N/156°04′W	79	–	2104	
						Kailua 20°54′N/156°13′W	213	–	3082	
						Paakea 20°49′N/156°07′W	384	–	5182	
Kauai (1598)	USA (Kauai, Hawaii)	22°06′N/159°30′W	MB shield caldera	4.0 Ma[c]	16–19 km	Lihue 21°59′N/159°21′W	103	24.0	1118	5
						Wainiha 22°12′N/159°34′W	101	–	3650	
						Waiahi 22°01′N/159°28′W	780	–	3650	
Tahiti Nui (2241)	France (Society Islands)	17°40′S/149°25′W	MB shield caldera	0.5 Ma	~8000	Papeete 17°32′S/149°35′W	2	26.0	1837[f]	12
						Hitiia 17°35′S/149°18′W	3	–	3152[f]	
Indian Ocean						Papenoo 17°31′S/149°25′W	2	–	3505[f]	
Piton des Neiges (3070)	France (La Réunion)	21°04′N/55°29′E	MB shield caldera	0.2 Ma[d]	~10,000[e]	Hellbourg 21°04′N/55°29′E	3070	17.0	2299[g]	13
						Sainte Rose 21°06′N/55°46′E	<5	–	3350[g]	
						Saint-Benoît 21°01′N/55°41′E	<5	–	4000[g]	

There may not be clear distinction between the two groups

[a] Maximum rainfall 1000–1100 mm at 1500 m altitude

[b–d] Youngest age of caldera collapse: [b] with Late Pleistocene to historic postcaldera activity; [c] with postshield- and rejuvenated-stage activity up to 0.52 Ma; [d] with an ultimate postcaldera activity of 0.03 Ma

[e] Average diameter of the three "cirques" (amphitheater valley heads); the whole, lobed depression has an 18-km average diameter

[f] Maximum rainfall 6000 mm in the caldera interior (1500 m altitude)

[g] Maximum rainfall over 6000 mm at moderately high altitudes in the east

Table 1b Topographic, morphometric, and climatic data for slightly eroded, breached craters and calderas around the world, with regions in ascending order of precipitation. Note the elevation differences between the highest points and the meteorological stations. Each additional 100-m increase in elevation corresponds to ca. –0.6 °C annual temperature drop and variably increasing (sometimes decreasing) rainfall, also controlled by exposure and orography

Volcano (highest point, m)	Country	Geographic position	Known or most likely origin of the present depression	Cessation (youngest age) of volcanism	Depression diameter (m)	Name and geographic position of the nearest meteorological station	Station elevation (m)	Annual temperature (°C)	Annual rainfall (mm)	Reference
The Carpathians										
Mohos (1245)	Romania	46°08′N/25°54′E	UB explosion crater in a dome complex	30–40 ka	1620	Sfintu Gheorghe 45°52′N/25°47′E	561	7.6	584[d]	1–3
North America										
Mt. Dana (1354)	USA (Alaska)	55°38′N/161°13′W	UB crater of a dome complex	3840 years BP	1900	Cold Bay 55°12 N/162°43′W	29	3.3	871	5
Okmok (1073)	USA (Alaska)	53°43′N/168°65′W	UB shield caldera	2400 years BP[a]	~9300	Cold Bay 55°12 N/162°43′W	29	3.3	871	5
Aniakchak (1341)	USA (Alaska)	56°53′N/158°10′W	UB stratovolcanic caldera	~3400 years BP[a]	~10,000	Cold Bay 55°12 N/162°43′W Kodiak 57°30′N/152°45′W	29 5	3.3 4.8	871 1380	5
Veniaminof (2507)	USA (Alaska)	56°11′N/159°23′W	MB(?) stratovolcanic caldera	~3700 years BP[b]	~9000	Cold Bay 55°12 N/162°43′W Kodiak 57°30′N/152°45′W	29 5	3.3 4.8	871 1380	5 5
Japan Islands										
Okamado-Yama (1150)	Japan (Kyushu)	32°52′N/131°03′E	UB stratovolcanic crater in the Aso caldera	?Late Pleistocene	650	Fukuoka 33°35′N/130°23′E Kagoshima 31°34′N/130°33′E	4 5	15.0 16.6	1592 2201	–
Aso (1236)	Japan (Kyushu)	32°30′–33°02′N/130°58′–131°41′E	UB caldera volcano	~70 ka[c]	18–24 km	Fukuoka 33°35′N/130°23′E Kagoshima 31°34′N/130°33′E	4 5	15.0 16.6	1592 2201	14
The Philippines										
Mt. Parker (1842)	The Philippines (Mindañao)	6°07′N/124°54′E	UB stratovolcanic caldera	1641 AD	2900	General Santos 6°07′N/125°12′E	<50	26.9	943	15

There may not be clear distinction between the two groups

[a–c] Youngest age of caldera collapse: [a]minor historic postcaldera activity; [b]ice-filled caldera with significant postcaldera activity; [c]Late Pleistocene to recent postcaldera activity; [c]Late Pleistocene to historic postcaldera activity

[d] Maximum rainfall 1000–1100 mm at 1500 m altitude

Selected references for both tables: 1 Szakács and Seghedi (1995); 2 Karátson (1996); 3 Karátson (1994); 4 Downes and Vaselli (1995); 5 Wood and Kienle (1990); 6 Cicacci et al. (1986); 7 Ollier (1988); 8 Moriya (1979); 9 Shimizu et al. (1988); 10 Simkin et al. (1981); 11 Macdonald (1978); 12 Bardintzeff et al. (1988); 13 Chevallier and Vatin-Perignon (1982); 14 Yanagi et al. (1993); 15 Delfin et al. (1997)

Sources of climatological data: World Survey of Climatology 1983, vols 1–15 (Elsevier, Amsterdam); Climate Normals for the US 1983 (Gale Research Company); The Weather Almanac 1987 (Gale Research Company); Climate of the Earth 1984 (in Hungarian; Textbook Publishing House of Budapest); “Le climat de la Polinésie Française” 1978 (“monographie no. 107 de la Météorologie Nationale”, France)

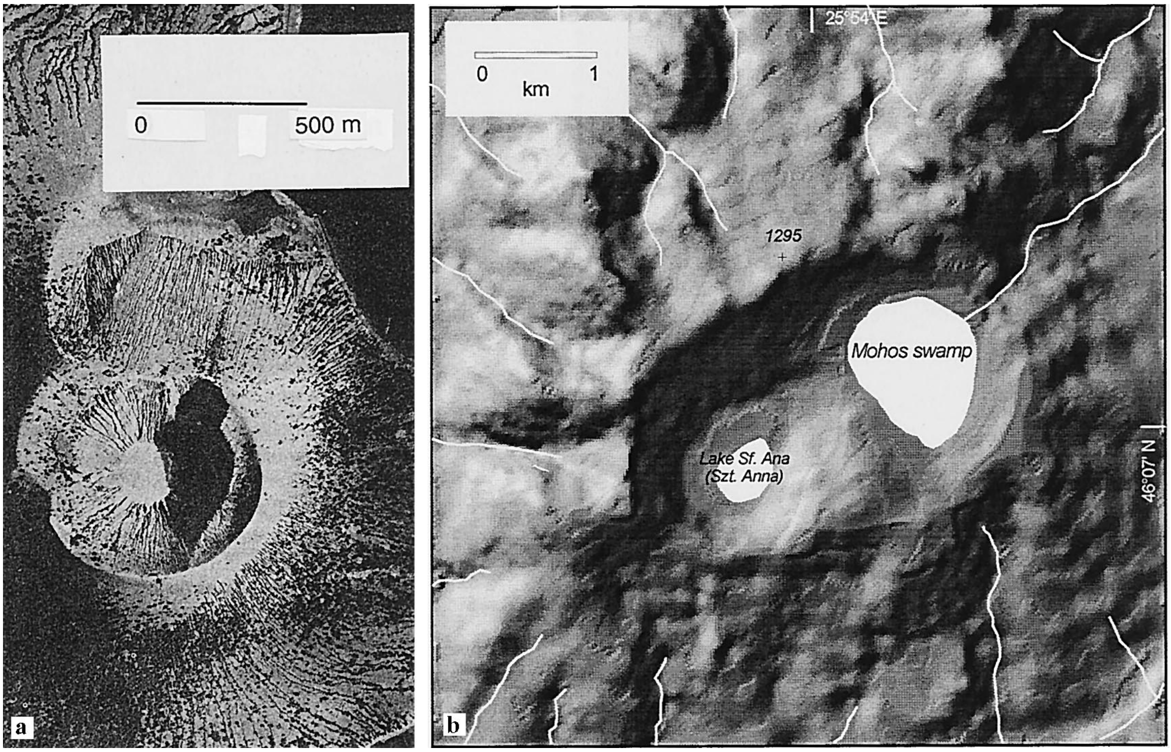


Fig. 5a, b First-stage crater degradation. **a** Aerial view of gullying on the active Vulcan pumice cone, Rabaul, New Britain, prior to the 1994 eruption (after Ollier 1988). **b** Computer-generated shaded relief map of the Late Pleistocene twin-cratered Ciomadul (Csomád) volcano, Harghita Mountains, Romania. The final activity of the unbreached Lake Sf. Ana (Szt. Anna) crater was 30–40 ka. Note the intense slumping on the south and east slopes. (Figures 5b, 6–10, and 14a–c were produced using digitized 1:50,000 topographic maps and AUTOCAD R14 and SURFER (Win32) 6.04 software.)

ippines (examples with 2000–3000 mm annual rainfall; Table 1). This suggests a precipitation threshold (ca. 2500 mm/year), below which different rates of erosion (such as much more rapid enlargement of the crater rim) do not change the style of degradation. The actual rim is derived from the original one.

Slightly and significantly eroded craters in Table 1 include examples (Fig. 6) from the Carpathians (Romania, Slovakia), Honshu and Kyushu (Japan), Luzon and Mindañao (the Philippines), and Alaska (U.S.). These regions are characterized by annual rainfall from 600 to 3300 mm. Consequently, climates below the proposed threshold range from semiarid through temperate to savanna and monsoon types. Examples from a given area, such as the Carpathians, with progressively enlarged crater diameters – Mt. Mohos (1621 m, 0.5 Ma), Ostoroş (Osztoróc, 2360 m, 6.3 Ma), Strechov (Sztrehó, 2720 m, 11.0 Ma; Karátson 1996) – show that unidirectional drainage can be maintained for millions of years, probably until the residual stage of degradation. In the Carpathians as well as in the more arid Cima field in California, morphometric changes have been uniform despite climatic fluctuations in the Pleistocene. Perhaps arid to moderate climatic conditions are less important than volcanological properties (material, volcano size, slope angle, etc.; Dohrenwend et al. 1986).

More intense, short-term endogenic and exogenic processes, such as faulting, may cause secondary or tertiary breaching. This may result in the formation of multi-drained depressions, typical examples of which are discussed later.

Erosion of closed summit calderas on basaltic shields or stratovolcanoes

Erosion-modified closed calderas are much scarcer than erosion craters because: (a) calderas, especially those of stratovolcanoes, commonly have a complex history resulting in different types of caldera opening or breaching; and (b) shield calderas are located mostly on oceanic islands that may experience large-scale landsliding (Duffield et al. 1982; Holcomb and Searle 1991) and/or intense rainfall, which occasionally causes multidirectional breaching (see below).

For the known continental examples (Table 1), degradation and drainage of closed summit calderas are similar to those of the craters mentioned previously. Figure 7 shows the drainage network of two calderas, Seaca-Tătarca (Mezőhavas, East Carpathians, Romania; Karátson 1994) and Elgon (East Africa, Kenya and Uganda; Ollier 1988), which are interpreted as breached by headward erosion. A "fixed" channel network and the dominance of one outlet valley are evident at Seaca-Tătarca, a shield-like edifice dominated

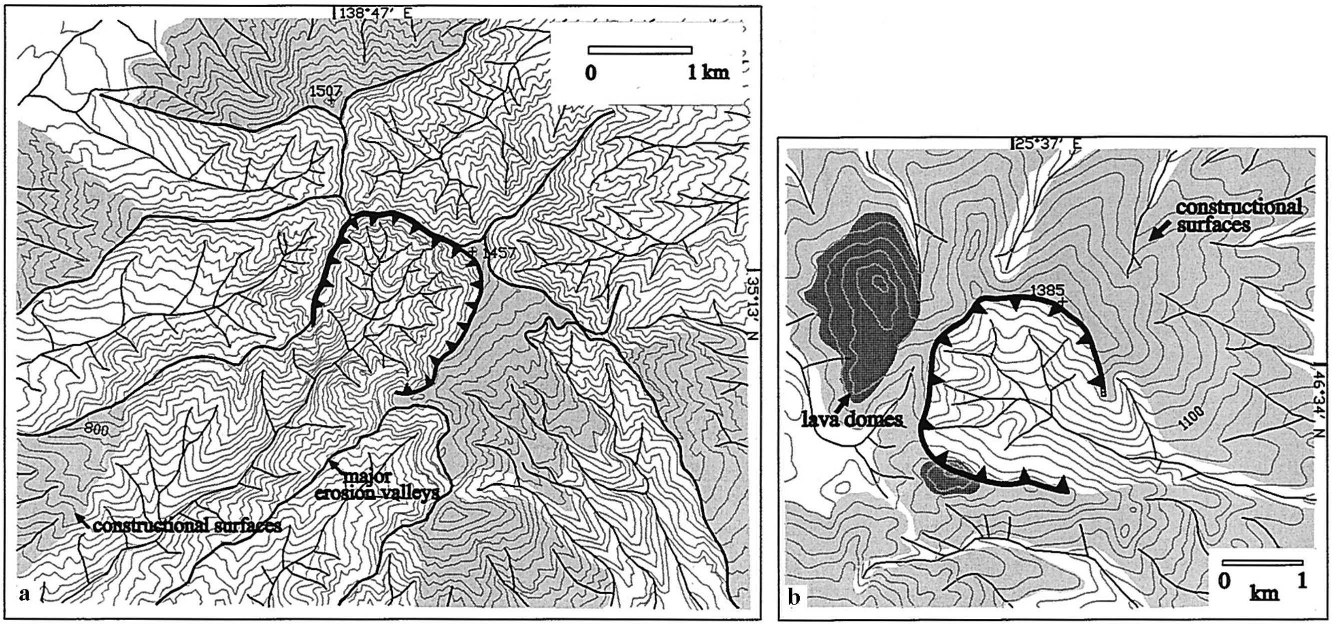


Fig. 6 Erosion craters: volcano-geomorphic sketch of **a** the Late Pleistocene Ashitaka volcano (Honshu, Japan), and **b** the Late Miocene Ostoroş (Osztoróc) volcano (East Carpathians, Romania), both with a unidirectionally breached crater (crater rim is indicated by *toothed line*; contour interval 40 m). Note that despite older age, constructional surfaces are better preserved on Ostoros volcano, owing to less annual rainfall

by basaltic andesite lava flows (Szakács and Seghedi 1995). Except for the outlet valley, none of the external valleys has breached the rim of Seaca-Tătarca caldera in the 7 million years since the last eruption.

The size of the depression suggests a preliminary distinction between erosion craters and erosion calderas, even if detailed geological information is lacking. Active craters are usually distinguished from calderas by having a diameter >1 km (Wood 1984). However, the largest recognizable erosion craters in Japan or in the Carpathians have diameters up to 2–3 km, whereas summit erosion calderas of the world usually exceed 3–4 km across (Table 1). This may indicate an approximately 3 km "erosion diameter" distinction between the two types. Consideration of other erosional features of a depression, beyond simply the diameter, is essential, however. Examples of such other features are the stage of crater degradation, the drainage density and stage, and the slope angles (Dohrenwend et al. 1986; Karátson 1996). These are particularly important because large craters and small calderas are common. Examples are large explosion craters of tuff rings, tuff cones and maars (up to 2–2.5 km in diameter on the Kamchatka peninsula; Braitseva et al. 1995), and relatively small, 1- to 3-km collapse or explosion calderas on stratovolcanoes (such as the early Holocene caldera of Mayor Island, New Zealand; Houghton et al. 1992; the middle to late Pleistocene Longonot caldera, Kenya; Buckle 1978; the 30-ka Ohachidaira caldera, Daisetsu volcano, Central Japan; Moriya 1979; and the 50-ka Akagi volcano, Central Japan; Moriya 1970). On the other hand, we emphasize that 2- to 3-km erosion craters, such as those in the Carpathians or in Central Japan, are unusual, because stratovolcanoes 1–3 km high can be eroded rapidly when exposed to extreme climatic processes, such as glaciation in the Cascade Range, (U.S.; Wood and Kienle 1990) and the northern part of Japan (Suzuki 1969).

Erosion of large caldera volcanoes

Low rims of calderas of large diameter (≥10 km; Moriya 1979) are commonly discontinuous, buried by post-caldera products, and/or cut by faults (Valles, Krakatau, Santorini, Cerro Galan, Aira caldera). An exception is Aso, South Japan (18–24 km in diameter), one of the largest calderas in the world, which is almost unbroken and drains only westward through the narrow, tectonically preformed Tateno gorge (Yanagi et al. 1993). Post-caldera activity and headward erosion on outer slopes have not breached the rim since 70 ka, the end of caldera formation. A similar scenario is likely for the Bracciano volcano-tectonic depression and Bolsena caldera, Italy. There, fault-determined outlet valleys (D. de Rita, pers. commun.) are the only drainage pathways.

Landslide depressions

On one level, depressions that form by large-scale landsliding can be considered as erosional features. Apart from the now-classic St. Helens type, there are examples of "pure" landslide-related depressions, mostly from oceanic islands. For example, there are well-docu-

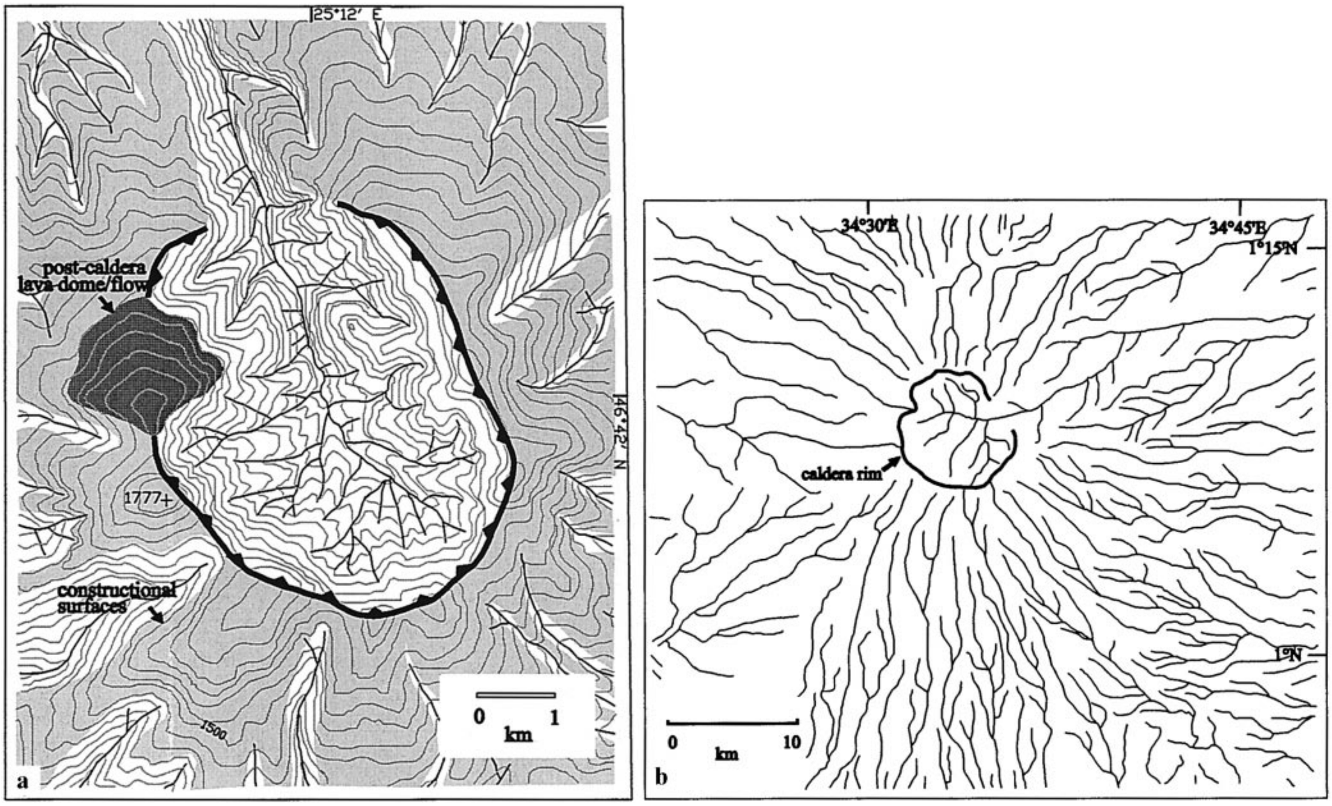


Fig. 7a, b Erosion calderas. **a** Volcano-geomorphic sketch of the Late Miocene Seaca-Tătarca (Mezőhavas) volcano (East Carpathians, Romania; caldera rim is indicated by *toothed line*; contour interval 40 m); **b** drainage network of the Pliocene Elgon volcano (Kenya/Uganda, East Africa) on the basis of the 1:250,000 scale topographic map of the Survey of Kenya

mented cases in the Canary Islands (Masson 1996), despite controversial hypotheses (Palacios 1994; Martí et al. 1996; Carracedo 1996). However, this process occurs on active rather than extinct volcanoes and is quite different from normal fluvial erosion. Furthermore, large-scale landsliding commonly results in open or half-open depressions that cannot be considered erosion calderas.

Nevertheless, there may be quasi-closed, oval-shaped depressions the formation of which is more or less related to small- and medium-scale landslides. One of these is the Tirajana Basin on Gran Canaria Island. It is a 35-km^2 flank depression with one outlet valley (Barranco de Tirajana) that is postulated to have been formed by repeated, mostly translational landslides from at least 600 to 125 ka (Lomoschitz and Corominas 1997). The present-day outlet valley is as narrow as those at other fluvially breached craters and calderas, perhaps because (a) post-slide fluvial erosion of Barranco de Tirajana has caused an apparent bottleneck morphology at the depression outlet, and (b) the Tirajana Basin is a multi-stage depression, and significant landsliding has occurred after the first-stage slides, enlarging the depression mostly in the upper portion of the valley. As the Tirajana depression is of modified landslide, not volcanic, origin, it should be grouped into the category of erosion-induced calderas, probably a common type of erosional depression formed in volcanic terrain.

Scarps or depressions in volcanic terrains resulting from landsliding are much more frequent morphological features than the unusual, quasi-closed Tirajana Basin; however, most often they are filled or masked by subsequent eruptions. Alternatively, sliding (and post-slide erosion) may result in different, open to quasi-closed primary depressions. The degree of "openness" could be a criterion for further subdivision.

Coalesced crater clusters

A specific type of erosion-induced caldera, formed under continental climate over a long time, has been described in the Carpathians (Fig. 8). For summit crater groups and clusters, the outer channel that has the highest valley-forming energy can breach all of the craters or capture their breaching channels, incorporating them into a shared, single-drained basin by normal fluvial processes [Morské Oko (Szinnai-kő), Vihorlat Mountains, Slovakia; Cucu (Kakukkhegy), Hargita Mountains, Romania; Karátson et al. 1992; Karátson 1999). Intense hydrothermal alteration, which commonly occurs in the crater areas above a single magma chamber, may promote development of such a depression. In contrast, multiple craters can be breached from

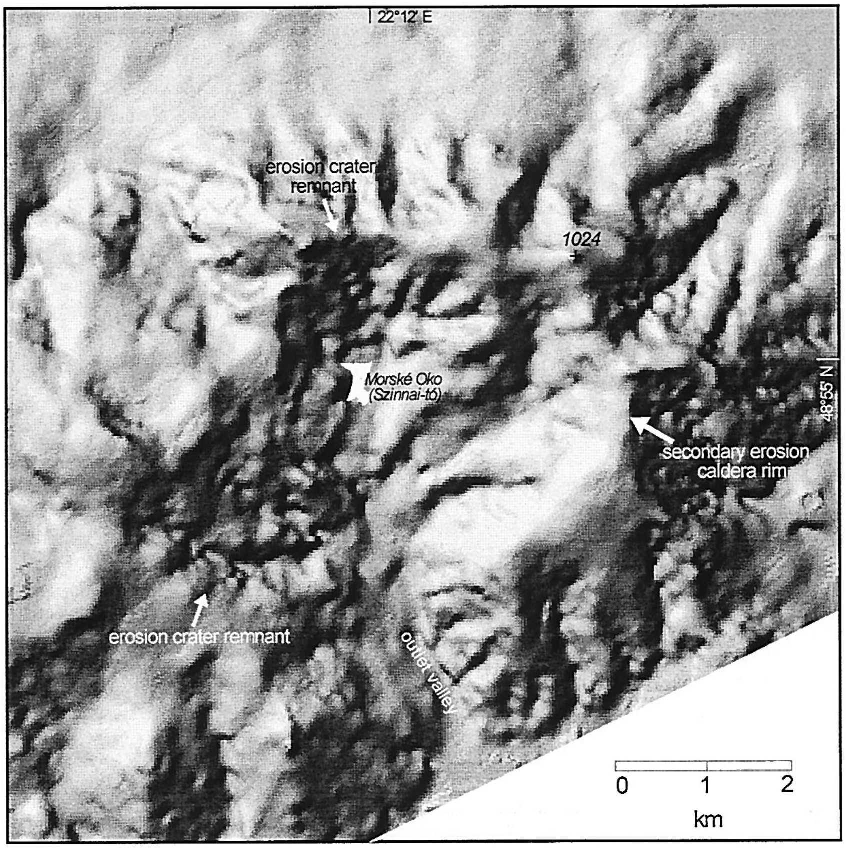


Fig. 8 Computer-generated shaded relief map of the Middle to Late Miocene Morské Oko (Szinnai-kő) volcanic edifice (Northwest Carpathians, Slovakia), showing two breached crater remnants in a caldera-like depression

opposite directions, if outer channels show similar erosive energy (Karátson et al. 1992). In such a case, each crater is further eroded by its own outlet valley without the formation of a shared basin (Fig. 9).

Depressions with several outlet valleys

Under heavy annual precipitation or intense rainstorms, primary volcanic landforms, such as craters, usually disappear over a geologically short time (Ollier 1988; Francis 1993); however, a special type of crater and caldera erosion sometimes results in surface features that can be preserved for a long time. This erosion process is related to *amphitheater valleys*, which are found mostly on low-latitude oceanic islands such as Hawai'i, Vanuatu, Tahiti, La Réunion, Mauritius, or parts of the Philippines. Instead of popular names, such as canyon, we use the term amphitheater valley after Hinds (1925) who first described them. We emphasize that the term is distinct from landslide-related amphitheater calderas.

Amphitheater valleys are large but relatively flat and have a cirque-shaped head (Figs. 10–13). They have a convex cross section, and the steep valley slopes are densely dissected into spurs. Studying such valleys in volcanic terrains, Cotton (1952) found that they develop under extremely heavy rainfall and that rapid tropical weathering is also essential. Indeed, "saucer-" or "bowl-shaped" valleys have been widely recognized in the humid tropics where annual rainfall exceeds 1000 mm (Faniran and Jeje 1983), but typical amphitheater valleys form when rainfall is at least 2000–2500 mm/year, and valley downcutting is commonly initiated or enhanced by rainstorms. Rainfall erosivity is positively related to rainfall intensity (Thomas 1994), and heavy rainstorms (often >200 mm/24 h) are common features of tropical weather between 20° N and 20° S (Thomas 1994). Occasionally, small-scale amphitheaters can develop under moderate rainfall (1500–2000 mm/year, such as at Chokai volcano, Central Japan). In Mindañao (Philippines), heavy rainfall (2500–3000 mm/year) and orographic conditions have resulted in development of moderate to large amphitheater valleys in the 2500-m-high, northeast-exposed ranges of the northern part of the island (Misamis Oriental and Occidental; Fig. 10; Table 1), whereas normal fluvial valleys typify the southern part (as at 1322-m-high Mt. Sibulan).

However, the formation of these valleys seems to be controlled by other factors in addition to climate-induced fluvial erosion. These factors occur mostly at ocean-island volcanoes. The most important is the existence of fault systems and rift zones typical of ocean-island volcanoes (Duffield et al. 1982; Fornari 1987; Carracedo 1994). There, due to steep slopes, fault scarps, and high-elevation differences (Table 1), propagation of amphitheater valleys is commonly guided by

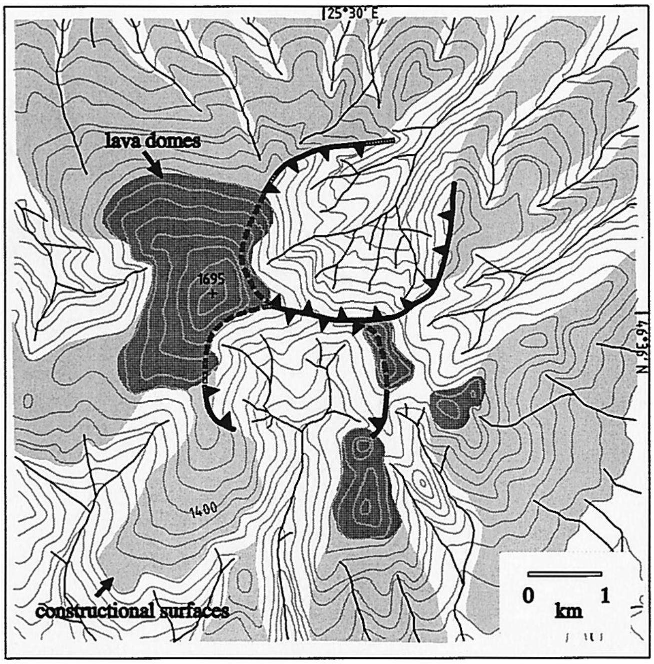
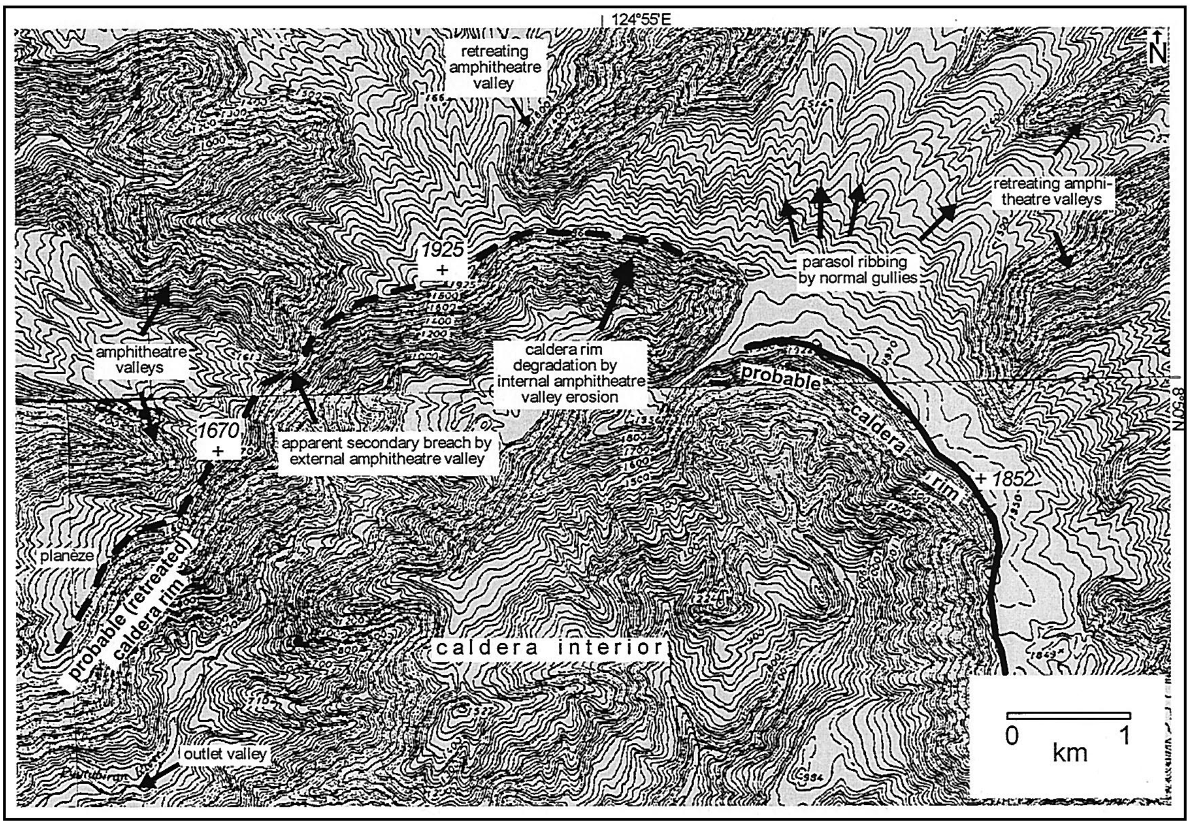


Fig. 9 Volcano-geomorphic sketch of the Late Miocene Ciumani (Csomafalvi Délhegy) volcano (East Carpathians, Romania) with two oppositely breached crater remnants. Crater rim is indicated by *toothed line*; contour interval 40 m

rockfalls and small-scale landsliding. The interbedding of resistant and non-resistant rocks, typical of both ocean-island and island-arc volcanoes, is also important (Ollier 1988). In less resistant layers (loose pyroclastic deposits or altered rock), valleys rapidly enlarge to an "amphitheater," whereas in the resistant rocks (such as lava flows), headward erosion, cascades, and waterfalls dominate. The contrasts in rock layers and related valley formation can result in "nested" or "multiple" valleys (Hinds 1925). Groundwater in ocean islands, which is determined partly by the heavy rainfall (Macdonald et al. 1983; Bachèlery et al. 1998), also contributes to high-rate alteration and erosion.

The size and erosion rate of amphitheater valleys well exceed those of normal fluvial valleys. At Haleakala volcano, East Maui (Hawai'i), valleys are several hundred meters deep, and the volcanic edifice is thought to be degraded by 700–1000 m (Harris and Tuttle 1990). On La Réunion Island (Indian Ocean), similar deep amphitheater valleys show a 2500-m vertical section in the Piton des Neiges volcano (Chevallier and Vatin-Perignon 1982). These valleys are youthful, <0.4 Ma in East Maui (Tilling et al. 1987; Langenheim and Clague 1987), <0.4 Ma in western Réunion (G.

Fig. 10 Detail of the 1:50,000 scale topographic map of Misamis Oriental (northern Mindañao, Philippines), revealing well-developed amphitheater valleys in and around the central caldera-like depression. Contour interval 20 m



Kieffer, pers. commun.), and <0.3 Ma in eastern Réunion (Mairine and Bachèlery 1997). The rate of downcutting is extremely high, up to 1 cm/year (Kieffer 1990; Mairine and Bachèlery 1997), especially if compared with stratovolcano degradation in continental climate.

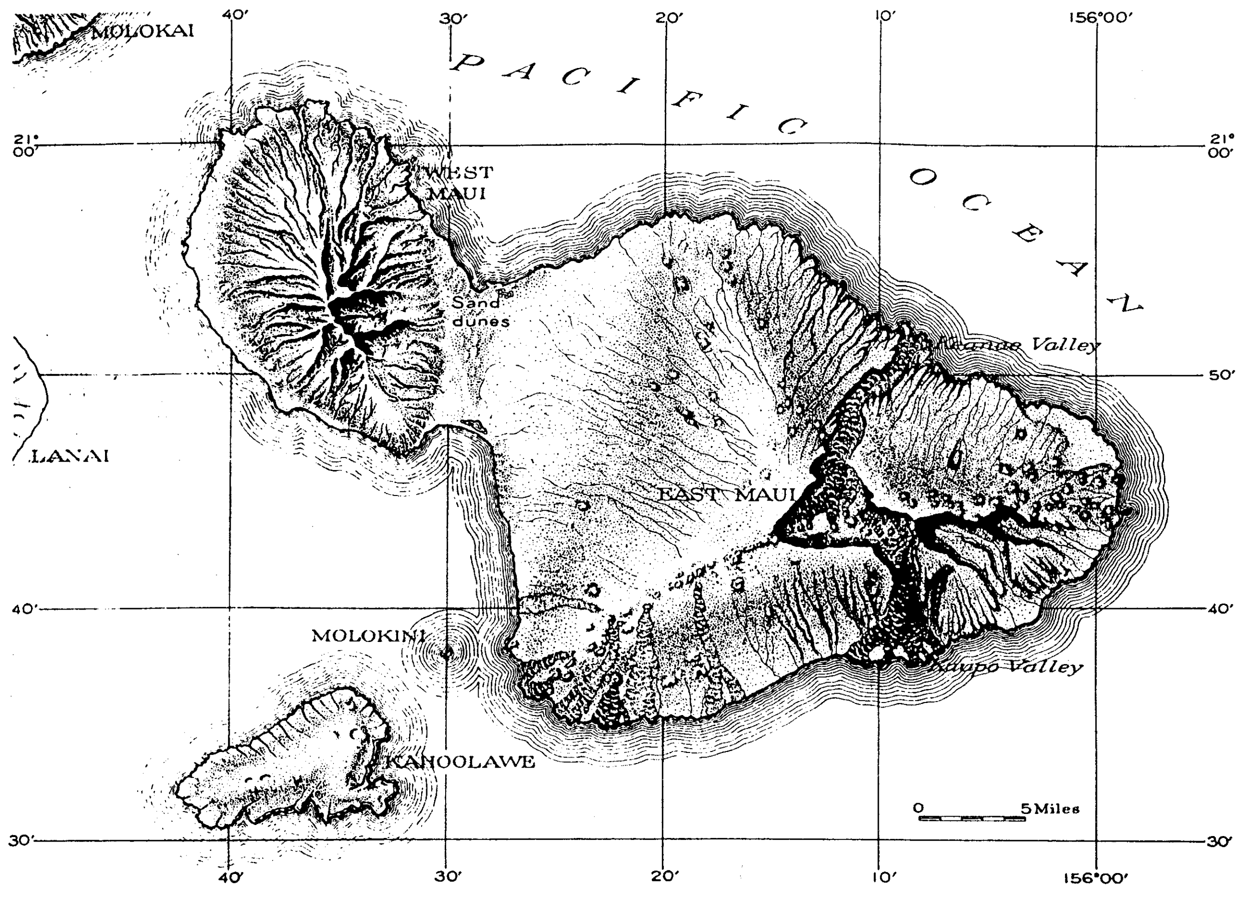
The extremely rapid formation of amphitheater valleys may explain some of their particular morphological features. Most importantly, several valleys can reach the summit area of the volcano and penetrate into a crater or caldera. In contrast to single valley erosion, such valley invasion results in rapid degradation of the summit area, in places exposing the subvolcanic rocks (as in the 0.5- to 3.5-Ma Tahiti Nui caldera, French Polynesia; Bardintzeff et al. 1988). However, such large amphitheater valley heads can survive as descendant landforms of the primary volcanic depression.

The principal types of multidirectionally breached volcanic depressions are described herein, with details and general rules of a few well-studied examples.

Erosion of calderas by amphitheater valleys

Haleakala Crater is an almost 50-km^2 depression in the summit of East Maui shield volcano (Fig. 11; Macdonald 1978; Tilling et al. 1987; Langenheim and Clague 1987). The depression has an uncommon, angular shape, with two large outlet valleys (Ke'anae, Kaupo) in opposite directions (north and southeast) and a third valley (Kupahulu) that has almost breached the rim. Dana (1849) explained its formation by fissure eruptions (also see the historical account by Appleman 1987). However, based on work by Stearns (1942) and Stearns and Macdonald (1942), it has been agreed that erosion of large amphitheater valleys was responsible for excavating the depression (Macdonald et al. 1983). The erosion took place after the formation of the 0.4-Ma-old Kula Volcanics, which may have built a 4000-m-high shield volcano, probably with a central caldera (Macdonald et al. 1983; Harris and Tuttle 1990). Easily removable pyroclastic deposits, altered lava flows, and groundwater supply may have contributed to rapid valley downcutting (Macdonald et al. 1983). Global glaciations elsewhere caused additional rainfall and frost action in the islands that resulted in extremely rapid headward erosion (Harris and Tuttle 1990). Amphitheater valleys (predecessors of the Ke'anae valley; see restoration map of Stearns 1942) could have breached

Fig. 11 Relief map of Maui island, Hawai'i. In western Maui, some controlling factors of amphitheater valley formation may have been lacking compared with eastern Maui, suggested by the fact that only the 'Iao valley has breached the inferred caldera. However, given the large amphitheater valleys around the caldera rim, valley coalescence in the summit region is likely in the future, as at Haleakala. (After Macdonald et al. 1983)



the crater starting from the north–northeast, where rainfall is heaviest (Table 1). Subsequently, the rim was breached from the south and southeast by other amphitheater valleys. Giant landslides on the southern flanks may have enhanced this process (Harris and Tuttle 1990). In contrast, erosion in the amphitheaters during the Late Pleistocene and Holocene was retarded by rejuvenated-stage lava flows and scoria cones, which partly filled the large valleys (Stearns 1942; Macdonald et al. 1983; Harris and Tuttle 1990).

On La Réunion Island, the active Piton de la Fournaise has moderately dissected slopes, whereas the inactive Piton des Neiges volcanic edifice is highly degraded; hence, different stages of erosion can be evaluated.

The relief of Piton des Neiges has a similar organization to that of Haleakala. Evolved amphitheater valleys with immense, circular heads up to 10 km in diameter, meet in its interior (Fig. 12). During the volcanic history (McDougall 1971; Chevallier and Vatin-Perignon 1982; Gillot and Nativel 1982), a submarine volcano evolved to a shield and filled a large initial caldera between 2.1 and 0.43 Ma. The end of this early stage may correspond to the initiation of amphitheater formation (G. Kieffer, pers. commun.). Secondly, a stratovolcanic center with a multi-stage summit caldera formed approximately 0.4–0.2 Ma (G. Kieffer, pers. commun.). The summit caldera was at the intersection of a radial fissure system, now revealed in the headwalls of the amphitheater valleys. From 0.35 to approximately 0.03 Ma (Gillot and Nativel 1982; Deniel et al. 1992; G. Kieffer, pers. commun.), the "differentiated series" of alkali lava flows and pyroclastic units was erupted and filled some older amphitheaters. During periods of volcanic quiescence, intense erosion controlled by fracture zones (Chevallier and Vatin-Perignon 1982) deepened three enormous amphitheater valleys (Cilaos, Mafate, and Salazie) far into the volcano (Upton and Wadsworth 1966). As at Haleakala, early erosion in the amphitheaters was intense and can be attributed partly to additional rainfall during humid periods of the Quaternary. In addition, the development of the amphitheaters was strongly related to lithological contrast. In the basal oceanite breccia of the first-stage, large-scale landsliding enlarged amphitheaters, whereas the upper series was more resistant to erosion (G. Kieffer, pers. commun.). Finally, the formation of large-scale amphitheaters resulted in multidirectional breaching of the summit caldera, obliterating the original cone and caldera shape and leaving planèzes in the summit region.

On the other hand, the original multi-stage summit caldera must have been a pre-existing depression that contributed to the enormous size of the present amphitheaters. This is clear if we consider that other, smaller

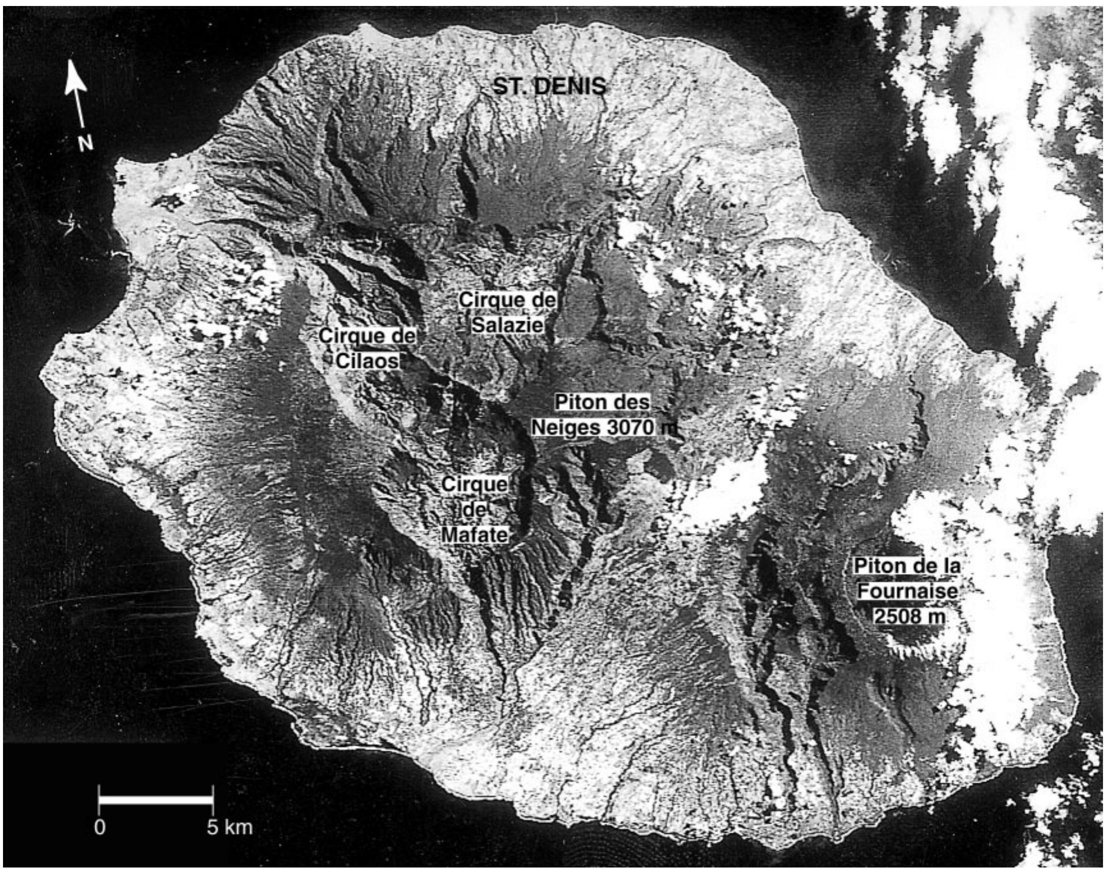


Fig. 12 LANDSAT satellite image of Réunion island

amphitheaters cut into the outer slopes of the large ones (Fig. 12). Furthermore, observations suggest a primary origin for at least the Cilaos amphitheater. That depression has surprisingly regular, circular contours with similar ridge height everywhere, and it has dimensions similar to those of Enclos Fouqué summit caldera in Piton de la Fournaise (ca. 8 km in diameter). Consequently, the original summit depression of Piton des Neiges may have been a multiple caldera, possibly with separate units. Given the subsequent intense erosion, the Cilaos, Salazie, and Mafate amphitheaters now almost coalesce; only a narrow, lowered lava ridge remains between them, and they may form a single but multi-drained basin in the future, similar to that of Haleakala.

A moderate stage of development, shape, and dimensions of amphitheater valleys can be observed on the active Piton de la Fournaise volcano, where erosion has not yet degraded the summit region (Fig. 13). In the central part of the volcanic edifice, "nested" caldera rims and scarps of huge landslide blocks can be reconstructed, together with the Enclos Fouqué caldera as an open-to-the-sea depression (McDougall 1971; Chevallier and Bachèlery 1981; Duffield et al. 1982; Gillot and Nativel 1989; Lénat 1990). Whether caldera collapses or flank landslides occurred, it is obvious that amphitheater valleys are structurally preformed; their asymmetric cross sections (Fig. 13) clearly show that they are connected to primary faults or scarps.

The intermediate stage of amphitheater valley erosion in Piton de la Fournaise is best seen around the Plaine des Sables area. The western rim, thought to be a caldera wall (Chevallier and Bachèlery 1981), is enlarged and excavated by typical amphitheater valleys of Rivière de l'Est in the north and Rivière des Remparts and Rivière Langevin in the southwest and south. New results show that a few enormous valleys evolved after an old shield-building stage, in part due to intense mudflows, approximately 0.3–0.2 Ma ago (Mairine and Bachèlery 1997). Similar to amphitheaters on Haleakala, subsequent lava flows filled the valleys and retarded erosion. Due to the young age, the amphitheater valleys, unlike those of Piton des Neiges, have not invaded the central part of the volcano, the <5000-ka-old Enclos Fouqué caldera; hence, Réunion amphitheaters show that several hundred thousand years are needed for such valleys to invade the center of a normal shield volcano.

In theory, long-term evolution (>0.5 Ma) of amphitheater valleys is best illustrated in the oldest Hawaiian Islands. However, many studies have shown that several factors control the morphology of the eroding shields in addition to development of "pure" amphitheater valleys: (a) caldera-filling stage more significant than at Haleakala; (b) subsidence of the islands and subsequent marine erosion; and (c) giant landslides related to subsidence, tilting, etc. Consequently, early recognition that amphitheater valleys coalesce to form lobed, retreating cliffs (as on O'ahu, Hawai'i; Cotton 1952; Wirthmann 1966) should be amplified by the role of giant landslides (Moore 1964; Fornari 1987; Holcomb and Searle 1991) and marine erosion (Wood 1990). Indeed, only one shield caldera, on Kaua'i island, shows features of the mature stage of amphitheater valley development, whereas most of the shields, such as Wai'anae and Ko'olau (O'ahu), Wailau (East Moloka'i), and Lana'i, have subsided calderas that are open to the sea and transformed by giant landslides. Dimensions of the enormous shield caldera of Kaua'i (16 × 19 km, Table 1) should obviously be attributed to long-term amphitheater valley erosion; moderate to large amphitheater valleys (the Waimea, Makaweli, and Wainiha "canyons") meet in the inferred caldera area. However, the caldera underwent a caldera-filling stage (Clague 1990), so the invading amphitheaters could not capture a large depression and deepen it greatly. As a consequence, the central area is presently occupied by the elevated, multi-drained Alaka'i swamp.

A mature stage of amphitheater valley evolution can also be observed in the Moka-Long Range on the island of Mauritius (Buckle 1978, Nohta et al. 1997); its highly degraded, steep ridges are the remnants of a Pliocene volcano.

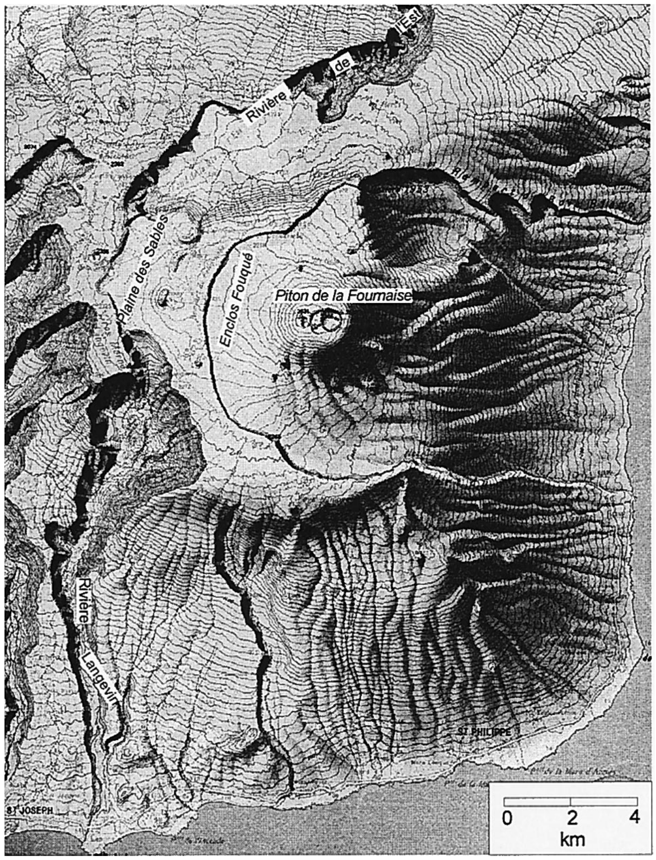


Fig. 13 Part of the shaded topographic map of Piton de la Fournaise volcano (from the Atlas of Institut Géographique National France 1956)

Formation of ocean-island depressions by amphitheater valleys without a caldera

We suggest that the high energy of amphitheater valleys alone, enhanced by frequent small-scale landsliding or wave erosion, may create an eroded internal lowland in the core of an ocean-island volcanic edifice that lacked a significant summit crater or caldera or the depression of which was filled during post-caldera activity. This may be the case in certain parts of Piton des Neiges volcano or in Efate, Vanuatu (Ollier 1988). In Kaho‘olawe island (Hawai‘i), excavation of the central filled caldera has been governed mostly by wave erosion (García 1990). The center of the western part of Tahiti was thought to be a dome excavated by amphitheater valleys (Cotton 1952); however, a recent study has pointed out a highly eroded shield volcano with a central caldera enlarged and deepened by amphitheater valleys (Table 1; Bardintzeff et al. 1988).

Multidirectionally breached depressions not related to amphitheater valleys

Several examples show that volcanic depressions with more than a single outlet valley can form in a climate without heavy rainfall if strong endogenic or exogenic factors are involved. Drainages of three of them are shown in Fig. 14. Makovica (Fig. 14a), one of the largest erosion craters in the Carpathians (Karátson 1996), has three small parallel outlet valleys; they may have been formed owing to intense faulting to the east (Kaliciak et al. 1986). Despite erosion of the three outlet valleys, the degradation of the crater is uniform and slow, as shown by the well-preserved shape. Luci (Nagykőbükk; Fig. 14b), a shield-like basaltic andesite volcano in the East Carpathians (Szakács and Seghedi 1995), has a medium-sized caldera with a flat interior occupied by a swamp. Its internal area presently drains westward, but a similarly evolved stream has cut down in the east at the caldera bottom (Karátson 1994). The low, interrupted rim of the depression, as well as gentle slopes around the volcano (related to the low-viscosity lava), enabled another channel to invade the caldera. The summit caldera of Akagi volcano, Central Japan (Fig. 14c), is presently drained by three valleys to the northwest, southwest, and southeast. The caldera contains two post-caldera lava domes on its floor (Moriya 1970). These elevated domes have created independent catchment areas that were breached and subsequently drained by different channels. Another similar example is the Late Pleistocene(?) Emmons Lake caldera, Alaska, the rim of which has been breached by post-caldera stratocones (Miller and Smith 1987). The caldera now has three outlet valleys and numerous rim notches due to strong glaciation.

Discussion and summary

Breaching and erosion of a crater or caldera can be initiated either by rapid volcanic and nonvolcanic processes (such as eruptions, geothermal activity, landslides, or debris flows) or by normal fluvial erosion. After eruptions end, the circular shape of the primary depressions can be preserved for millions of years, although fluvial erosion results in a significant change in the drainage network development; therefore, we restrict the term "erosion crater or caldera" to *breached, fluvially eroded depressions of extinct volcanoes*. Landslide- or glacier-transformed volcanic depressions have more open "breaches" and strongly modified shape, and are not considered to be erosion calderas.

Climate is one of the most important factors in determining the number of fluvially eroded outlet valleys and therefore the style of depression degradation. To demonstrate this quantitatively, depressions of the same or similar origin – erosion calderas of composite and shield volcanoes (Table 1a) – are compared in Fig. 15, plotting age vs rainfall vs depression axis ratio (shortest/longest diameter). Figure 15 shows that unidirectionally and multidirectionally drained depressions are clearly separated by an annual rainfall of 2000–3000 mm. Below this threshold are calderas affected by normal fluvial erosion; above the threshold are more transformed depressions degraded by amphitheater valleys. When evaluating climatic vs structural control as factors of amphitheater valley erosion, it is of great importance that amphitheater valleys evolve not only in ocean-island shields but also in island arc stratovolcanoes (Fig. 10). Consequently, structural control (mostly that of ocean islands) may only supplement the impact of heavy and intense rainfall on the formation of amphitheater valleys.

Another, possibly genetic, implication of the plot is the difference in depression axis ratios. Unidirectionally drained depressions tend to be elongate (in response to erosion of a single outlet valley) and characterized by a relatively small axis ratio (0.6–0.8), whereas multidrained depressions are eroded by more valleys from several directions, so their axis ratio is greater, commonly >0.8. Haleakala "Crater,"an exception, may exemplify a structurally controlled depression, transformed by amphitheater valleys that were originally offset (Stearns 1942).

Craters and calderas with little or intermediate annual rainfall (up to ~2000–2500 mm) are commonly eroded by a single outlet valley formed by enlargement of the first breach. The outlet valley maintains its dominant downcutting position, because it rapidly increases its drainage basin by capturing the crater or caldera area. Such a drainage may be fixed until the residual stage of degradation, because other channels with smaller catchment areas on the outer volcano slopes are unlikely to incise at a faster rate and thus are unlikely to capture any of the crater or caldera area. How-

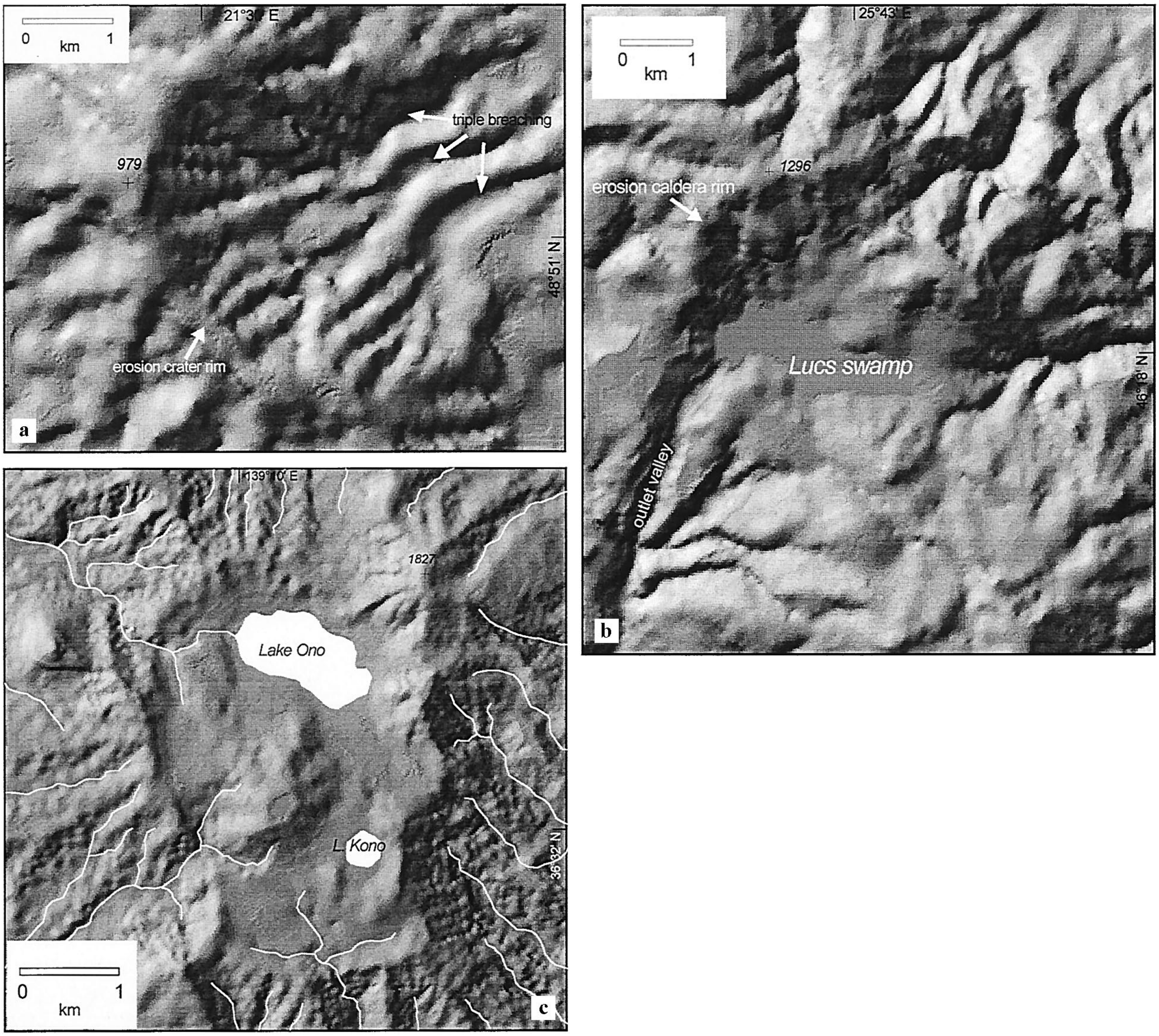


Fig. 14a–c Multiple breaching due to special morphology. **a** Computer-generated shaded relief map of the Late Miocene Mt. Makovica erosion crater (Northwest Carpathians, Slovakia). **b** Computer-generated shaded relief map of the Early Pliocene Luci (Nagykőbükk) erosion caldera (East Carpathians, Romania). **c** Computer-generated shaded relief map of the Late Pleistocene Akagi erosion caldera (Central Japan), completed with drainage network

ever, secondary breaching can occur if high-rate geological processes, such as faulting, suppress fluvial erosion. If so, a multi-drained depression can form, although the intensity of fluvial erosion may not change significantly. This means that the crater or caldera origin of the primary depressions is commonly recognizable; hence, they should be termed *erosion-modified volcanic depressions*. Exceptionally, the final negative landform is not derived from a primary volcanic depression. These *erosion-induced depressions in volcanic terrains* include crater clusters that are involved in a shared, single-drained basin, and quasi-closed flank depressions formed by landslides and fluvial erosion.

In climates with heavy, concentrated rainfall exceeding ~2000–2500 mm/year, multidirectional drainage is a much more frequent, if not normal, feature of crater and caldera erosion. In this case multiple drainage is related to amphitheater valleys. These valleys develop most often on low-latitude ocean-island or island-arc volcanoes with high precipitation. Several such high-energy valleys can penetrate into a closed crater or caldera, especially if aided by structural, hydrogeological, and topographic factors, and are propagated by intense mass wasting, landslides, and mudflows, thereby degrading the original depression from different directions. Some examples suggest that internal depressions can also form merely by the propagation of amphi-

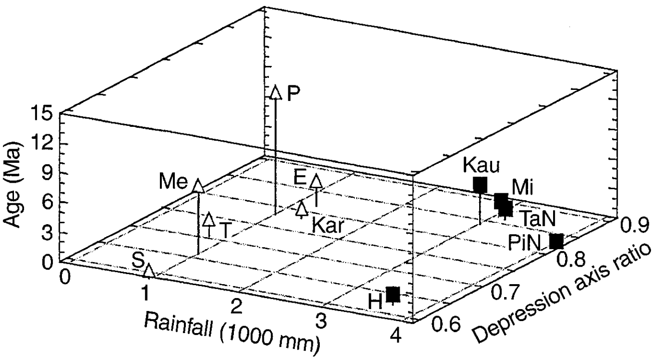


Fig. 15 Plot of rainfall vs age vs depression axis ratio of erosion calderas of composite and shield volcanoes. Rainfall values are averaged from Table 1. *Depression axis ratio* is ratio of shortest and longest caldera diameters. *Open triangles* are unidirectionally breached depressions, *black quadrangles* are multidirectionally breached. *S* Sacrofano; *Me* Seaca-Tătarca (Mezőhavas); *T* Taylor; *P* Pol'ana (Polyána); *E* Elgon; *Kar* Kariba; *H* Haleakala; *Kau* Kauai; *Mi* Misamis Oriental, *TaN* Tahiti Nui; *PiN* Piton des Neiges

theater valleys into the summit of an ocean-island volcano with an insignificant or filled caldera. Given the high-rate erosion of amphitheater valleys, multidrained "erosion calderas" in high-precipitation volcanic terrains should be termed as *erosion-transformed volcanic depressions,* indicating that the original negative landform is masked to such an extent that the preexisting crater or caldera, if any, can barely be inferred.

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