Geological Hazards in the Teide Volcanic Complex

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Abstract

The island of Tenerife displays contrasted densities of population, from the densely occupied coastal zones (including tourist resorts, airport, energy facilities, etc.) to the sparsely populated forests and mountainous highlands, where most of the recent volcanic events are located. Considering the low frequency of historical eruptions (compared to Hawaii or Reunion Island for example), the assessment of geological hazards must also rely on the analysis and interpretation of prehistorical events, going back to at least the Late Quaternary. In this chapter, we review the hazards related to Teide's volcanism, but also those from increased seismicity and from slope instability. We discuss the origin of low magnitude earthquakes, and particularly the 2004 episode of unrest. New estimates on cumulative volumes for resurfacing by lava flows during the last few thousand years are provided to serve as a tool for building a lava flow hazard map of Tenerife. Hazards related to explosive

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activity are also considered and although possible, with phreatomagmatic eruptions being the most likely style anticipated, explosive events are of relatively low probability at Teide in the near future.

14.1 Introduction

Geological hazards are comparatively moderate in the Canary Islands in general, and in particular in the TVC. Rock falls, road blocks, coastal cliff slumps and *barranco* flash floods are, by far, the main cause of loss of life and property, mainly at the lower part on the northern, leeward side of the TVC (Fig. 14.1).

This chapter analyses the most characteristic and significant hazards associated with the TVC as an active volcanic complex, namely seismic hazards, eruptive hazards and those derived from ground deformation.

In an early and very influential geophysical study on the Canaries, Dash and Bosshard (1969) and Bosshard and MacFarlane (1970) postulated large, regional "crustal fractures" related to the African tectonics that apparently pre-date the formation of the Canaries. This seminal idea opened a long-lived debate not only on the origin of the Canaries, but on the potential risk of major "tectonic" earthquakes in the archipelago.

Geophysical and geological information increasingly favours the construction of the Canary Islands in relation to a mantle plume on a passive continental margin disassociating the Canaries from northern African seismicity and tectonism (Frizon de Lamotte et al. 2000; Martínez and Buitrago 2002; Faccenna et al. 2004) (Fig. 14.2). The intense and abundant seismicity concentrated in the plate boundary zone between the converging African and Eurasian plates fades in the oceanic prolongation of the Atlas fault system, creating a seismic gap between Africa and the Canaries (Fig. 14.3). This reflects the difficulties encountered by tectonic stresses to propagate from the weaker continental to the stronger oceanic crust (Steckler and Tenbrink 1986).

Fig. 14.1 Rockfall and associated road blockage on a road on the northern slopes of the TVC (July 5, 2011, www.laopinion.es)

Geological hazards in the Canaries are thus correlated with a separate geodynamic framework, characterised by low magnitude seismicity (only 1 event $M > 5$ in the recording period), and volcanic eruptions where the majority of Holocene eruptions have $VEI < 3$.

In addition to the favourable geological and geodynamic setting of the Canary Islands in relation to natural hazards, passive margins around the Atlantic Ocean are characteristically aseismic, which contrasts, e.g., with the Hawaiian Islands. Consequently, the Canaries are at low risk of significant tsunamis induced by large earthquakes, which in turn frequently reach the Hawaiian Islands (Wood et al. 2007).

By comparing the islands of Tenerife and Hawaii (the classical example of an age progressive island chain fed from a hotspot source fixed in the mantle), it becomes apparent that the larger volume and more active mantle plume of Hawaii explains the considerably greater severity of geological hazards there, at least in the historical period of both archipelagos.

14.2 Seismicity and Seismic Hazards in the TVC

Although intraplate earthquakes do occur in some continental lithospheric plates (e.g., North American, Eurasian), seismicity is generally

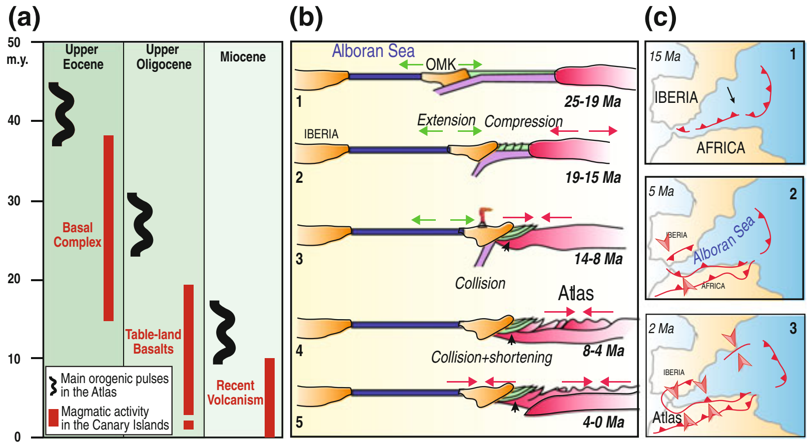

Fig. 14.2 **a** Correlation established by Anguita and Hernán (1975) between the main phases of Canarian volcanism and the three orogenic pulses involved in the genesis of the Western High Atlas structures (Ambroggi 1963). **b** Cenozoic evolution and active deformation in the North African margin (from Frizon de Lamotte et al. (2000). **c** Atlas folding apparently began in Early Pleistocene (from Faccenna et al. 2004)

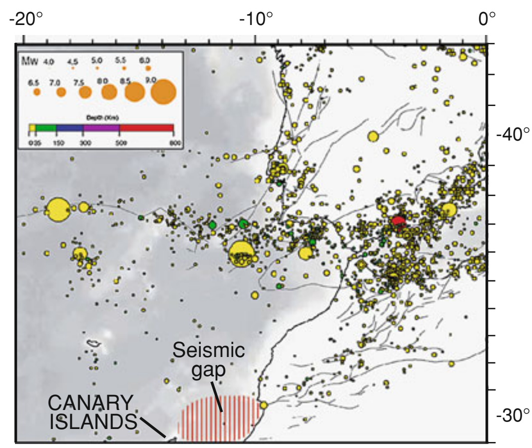

Fig. 14.3 The abundant seismicity in North Africa fades in the oceanic prolongation of the Atlas towards the Canary Islands, creating a seismic gap that apparently contradicts the alleged link of the archipelago with the Atlas tectonics. The figure shows earthquakes with M > 4 recorded from 1901 to 2004 (modified from Serpelloni et al. 2007)

negligible in the interiors of the lithospheric plates without the presence of oceanic islands (e.g. Hawaiian and Canary Islands in Fig. 14.4). Hawaii, for example, is far removed from geological fault lines, but the hot spot below the Hawaiian Islands causes sufficient stress in the crust to render Hawaii a very active seismic region (e.g., Klein et al. 2001). Earthquakes in these settings are thus mainly generated by plume related magmatism and the stresses caused by island growth and associated lithospheric loading.

Some authors, following the early ideas of Dash and Bosshard (1969) and Bosshard and MacFarlane (1970) relate seismicity in the Canaries with regional crustal fractures or to the Atlas system (Anguita and Hernan 1975, 2000; Geyer and Martí 2010; Mezcua et al. 1992). However, no compelling evidence has been found to support the existence of any major fault connecting the Atlas with the Canaries in detailed geophysical studies of the area (Martínez and Buitrago 2002) or around the Canarian archipelago in general (Watts 1994, 1997; Funck et al. 1996; Urgeles et al. 1998; Krastel et al. 2001; Krastel and Schmincke 2002; Carracedo et al. 2011).

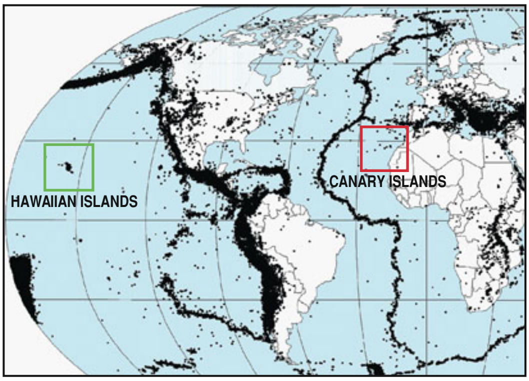

Fig. 14.4 M > 3.5 earthquakes (200,855 events) registered in the period 1963–1998. Seismicity in the interior of the plates is mainly related to the construction of the oceanic islands (from Earthquake Seismicity Catalog (2012), NOAA, National Geophysical Data Center, World Data Center-A Solid Earth Geophysics. http://www.ngdc.noaa.gov/hazard/earthqk.shtml)

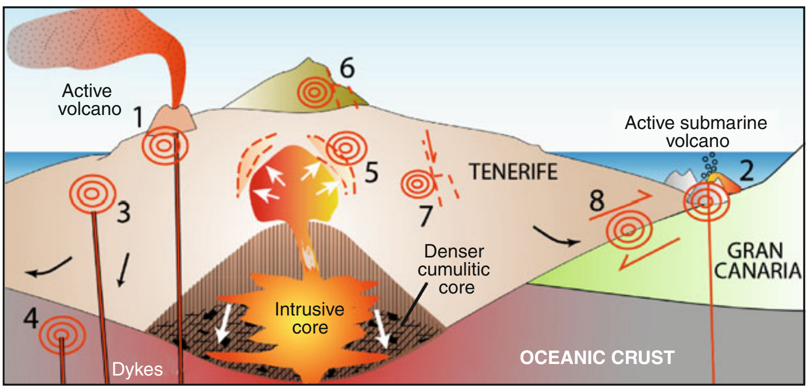

Fig. 14.5 Different types of earthquakes in oceanic islands: *1–2* earthquakes caused by volcanic eruptions (submarine or subaerial), including harmonic tremor generated in the eruptive conduits; *3–4* seismicity related to hydraulic fracturing by dykes in the crust or in the island edifice, that seldom reach the surface to produce a volcanic eruption; *5* earthquakes caused by inflation of shallow magmatic chambers; *6–7* earthquakes due to gravitational normal faulting; *8*. earthquakes associated with low-angle, inverse faults that originate by gravity-driven lateral escape of island sectors in the denser island basement, e.g. Volcano spreading (Carracedo et al. 2005). See text for explanation

In his work on the Andes, von Humboldt affirmed that "the large destructive earthquakes have no direct connection with volcanic activity, that only cause small local seismicity…" (von Humboldt and Bonpland 1805). Klein et al. (1987) deduced from studies in Hawaii that "All earthquakes in Hawaii are ultimately attributable to volcanism…", defining those earthquakes as "volcanic" that are in or immediately adjacent to magma conduits, and "tectonic" earthquakes as those that are removed from a magma conduit but derive their driving stress from dyke intrusion or volcano growth and deformation. In agreement with these concepts, seismic hazards decrease in the Canarian Archipelago with island age, as do both magmatic activity and edifice instability. Earthquakes in oceanic islands thus have different causes that group in two main categories: those associated with magmatic processes, and those originated by instability in the islands (Fig. 14.5). Earthquakes caused by volcanic eruptions (submarine or

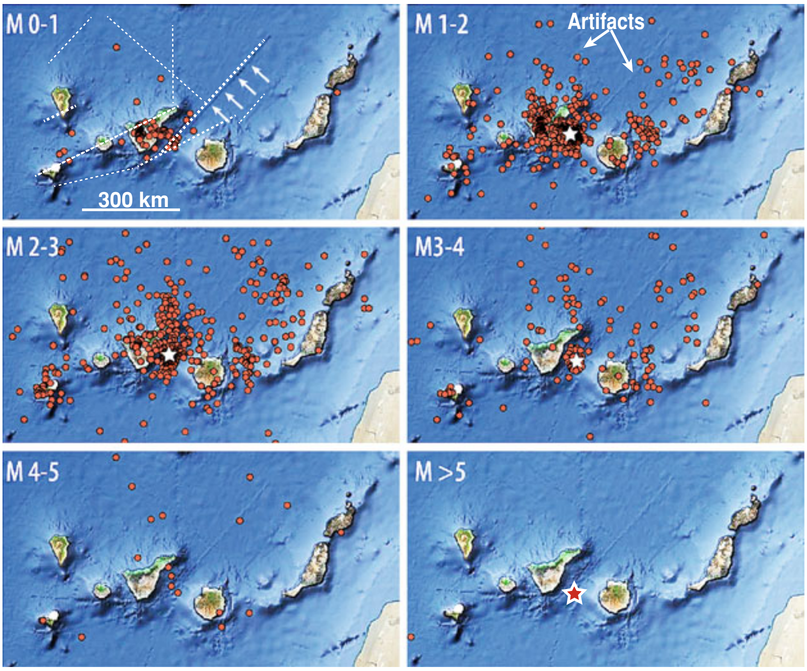

Fig. 14.6 Earthquakes recorded by the Spanish Geographical Institute (IGN) in the period 1980–2010, arranged as a function of their magnitude. The greater part of this seismicity has magnitudes below M 3.0. Only one earthquake of $M > 5$ has been recorded in the Canaries (M 5.2, May 1989). The epicentre distribution lacks any concordance with linear structures, such as the alleged faults of Geyer and Martí (2010) (*white dashed lines*), which are actually artifacts from *ship track lines* (Google Earth image). *The asterisk shows* the location of an active submarine volcano (Krastel and Schmincke 2002)

subaerial) belong to the first group and include harmonic tremor generated in the eruptive conduits (1 and 2 in Fig. 14.5). This category also includes the seismicity related to hydraulic fracturing by dykes in the crust or within the island edifice, that seldom reach the surface to produce a volcanic eruption (3 and 4 in Fig. 14.5); or that caused by inflation of shallow magmatic chambers (5 in Fig. 14.5).

The second category comprises earthquakes due to gravitational normal faulting (6 and 7 in Fig. 14.5), or associated to low-angle, inverse faults that originate by gravity-driven lateral escape of island sectors in the denser island basement, e.g. volcano spreading (8 in Fig. 14.5).

The majority of earthquakes in the Canaries in the period 1980–2010 (Fig. 14.6) have magnitudes between M1 and M3. The greater part of this seismicity is concentrated in the area of Tenerife, with two apparently different sources, one focused between Tenerife and Gran Canaria, and the other in the western part of Tenerife. The former is clearly defined by M 3–4 events forming a tight cluster, whereas the latter group is more disperse and shows only magnitudes $M < 3$.

The group of epicentres between Tenerife and Gran Canaria coincides with the alleged fault of Dash and Bosshard (1969) (dashed lines in Fig 14.6) and with a group of submarine volcanoes, some of them (e.g., Hijo de Tenerife) likely active (Schmincke and Rihm 1994; Schmincke and Graf 2000). As shown in Figs. 14.6 and 14.7, the epicentres between Tenerife and Gran Canaria present a Gaussian rather than the linear distribution which would be expected in case of fault-related seismicity (Carracedo et al. 2011).

The seismicity recorded for the island of Tenerife is comparatively low, particularly if compared with the island of Hawaii for a similar period of time (Fig. 14.7).

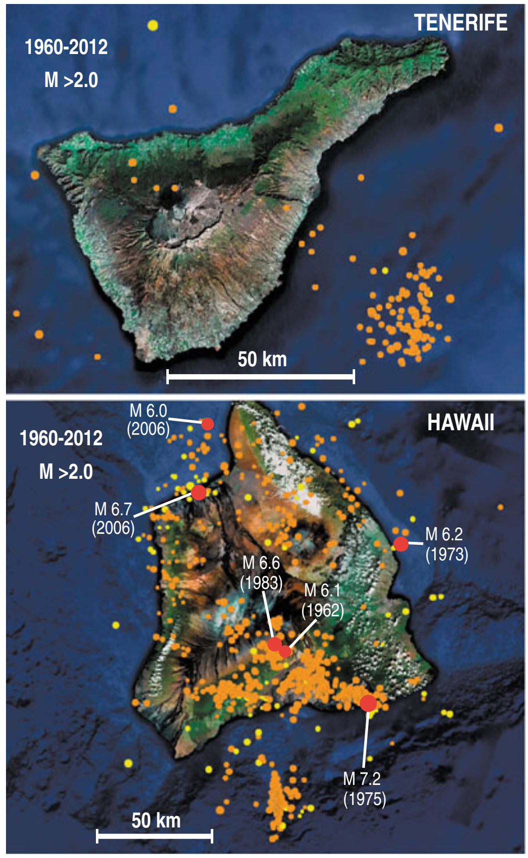

Fig. 14.7 Comparison of the M > 2.0 seismicity of Tenerife and Hawaii in the period 1960–2012 (IGN and USGS Earthquake Hazards Program, Rectangular Area Earthquake Search http://earthquake.usgs.gov/earthquakes/eqarchives/epic/epic_rect.php)

In the Hawaiian Islands, where high-quality earthquake hypocentre data have been collected by the Hawaiian Volcano Observatory since the late 1950s (the Hawaiian Volcano Observatory (HVO) has collected earthquake data since 1912, but its seismic network was modernised in the late 1950s). A wealth of information on seismic and magmatic processes is available from over 60 stations on the Big Island set up to monitor volcanic and tectonic features. In the Canaries, the period of adequate instrumental seismic data collection is very short and the number of seismic stations scant. The analysis of seismicity and assessment of seismic hazards associated with the TVC, in turn, has been hindered by the lack of instrumentation and, consequently, a database capable of establishing a baseline to define potential unrest episodes. The increase in the number of low magnitude events (M < 2.0) in May 2004, with several felt in the NW of the island according to IGN reports (M 2.0 to M 2.6, yellow dots in Fig. 14.8), was considered by some authors to be unusual "with much evidence pointing to a reawakening of volcanic activity" (García et al. 2006; Gottsmann et al. 2006; Martí et al. 2009; Dominguez-Cerdeña et al. 2011). However, other authors contested this alarming claim based on the lack of baseline data to detect and interpret which movements are normal for the area and which might indicate volcanic unrest (Carracedo et al. 2005; Carracedo and Troll 2006).

Notably, this apparently "anomalous" seismic activity coincided with the upgrade of the seismic network of the island (one station located in the northeast of the island and one inside the Las Cañadas Caldera). After the 2004 alarm—that ended without any observed volcanic manifestation but had a strong negative impact in the local and international media—the

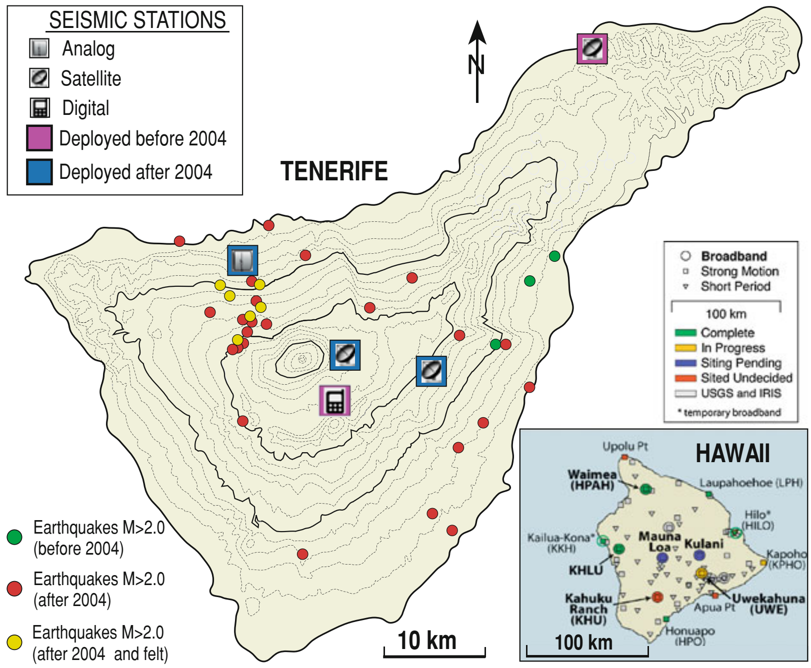

Fig. 14.8 Seismic network on Tenerife by the Spanish Geographical Institute (IGN) and $M > 2.0$ earthquakes recorded both before and after 2004. The seismic network deployed on the island of Hawaii is shown in the inset for comparison (from Shiro et al. 2009)

seismic network was upgraded to the present configuration of 5 stations, 4 of them deployed around the TVC (Fig. 14.8).

The origin of these earthquakes sparked much controversy within the scientific community. Although a deep magmatic intrusion was the favoured hypothesis, several different models were proposed. Using InSAR and GPS networks, Fernandez et al. (2003) detected deformation in the NW part of Tenerife, interpreted as subsidence caused by groundwater pumping operations performed on the island. These authors later associated the seismic crisis in 2004 to localized deformation correlated with the observed water table changes (Fernandez et al. 2005).

Gottsmann et al. (2006) observed gravity changes in the central part of Tenerife, relating the 2004 reactivation to a sub-surface mass addition without any significant widespread surface deformation, most likely caused by fluid migration at depth (arrival of new magma at depth, migration of hydrothermal fluids, or a combination of both).

Martí et al. (2009) regarded the unusual behaviour starting in spring 2004 as unequivocal evidence of volcanic unrest, based on the significant increase in seismicity onshore, the presence of volcanic tremor, the perturbation of the gravity field, the increase in fumarolic activity at Teide's crater and the opening of a new fracture with gas emission in the Orotava Valley.

Although a deep intrusion is a plausible explanation as to the origin of this perturbation, the alleged signs pointing to more specific volcanic unrest (e.g. harmonic tremor and anomalous gas emissions in the summit crater and inside the Orotava Valley) are far from evident.

The 2011/2012 volcanic eruption at El Hierro, with initially similar amplitudes, sheds new light on the 2004 Tenerife unrest. Numerous earthquakes M < 3.0 were recorded north of El Hierro by the Spanish Instituto Geográfico Nacional (IGN) since July 2011 related with intrusion of magma within the lower oceanic crust. However, conversely to what occurred in

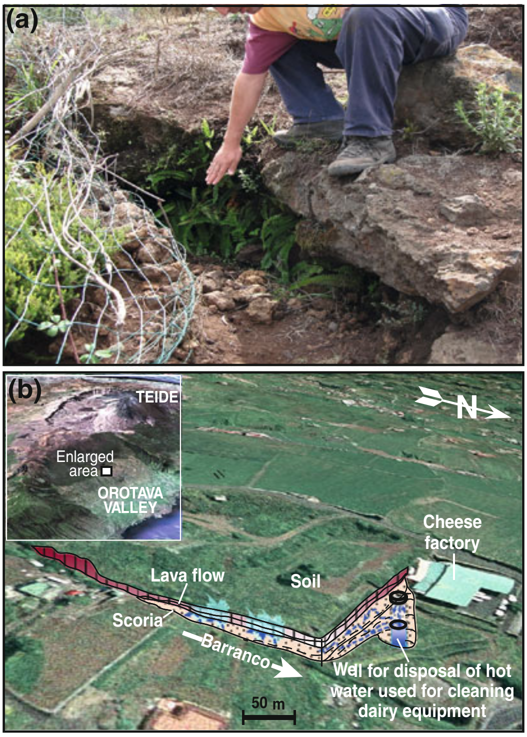

Fig. 14.9 **a** Vent with vapour emissions with temperatures of 15–30 °C in the Orotava Valley, interpreted as a new fumarole in 2004 related to a N–S trending fracture. **b** Geological cross-section illustrating another explanation—the residual vapour emissions of hot water used for cleaning the equipment of a cheese factory located 80 m distant (after Carracedo and Troll 2006)

2004 in Tenerife, shifting seismic foci suggested that magma progressively accumulated and expanded laterally in a southward direction, causing a vertical surface deformation of about 40 mm (data from IGN). In October and November earthquake magnitude increased significantly (M > 3.0), with 5 events of M > 4.3 (IGN). The scenario changed dramatically at about 4 am on October 10, 2011, when the now frequent and strong seismicity ceased and was rather abruptly replaced by a continuous harmonic tremor, indicating the opening of a vent and thus the onset of the submarine eruption (Carracedo et al. 2012a, 2012b). A similar process of deep magma intrusion probably started in 2004 in Tenerife, but halted before approaching shallow levels to induce strong seismicity, ground deformation and eruption.

Almendros et al. (2000) carried out a field experiment in the TVC in 1994 with two short-period small-aperture seismic antennas. The objective of the experiment was to detect, evaluate and locate the local seismicity, particularly the background seismic noise to investigate the possible presence of volcanic tremor. From their work they concluded that at frequencies lower than 2 Hz, the oceanic load signal was predominant over other signals, consequently masking any weak volcanic tremor or other volcanic signals with predominant peaks below 2 Hz present in the Teide area.

Ten years later Almendros et al. (2007) repeated a similar experiment (with three seismic antennas deployed in the Las Cañadas Caldera) during the most active period of seismicity in the area (May–July 2004). This time they detected a continuous tremor (small amplitude, narrow-band signal with central frequency in the range 1–6 Hz) that started on May 18 and lasted for several weeks. This was the first time that volcanic tremor had been reported at Teide volcano. The model these authors outlined to explain the origin of the volcanic tremor is based on a deep magma injection under the northwest flank of Teide volcano. The primary process driving the tremor would be related to interactions between a shallow aquifer and an increased discharge of magmatic gases supplied by deep magma injection.

Finally, the new fracture with gas emission claimed to have opened in December in the Orotava Valley (Martí et al. 2009) is not a fracture but a lava flow cut by a *barranco* (Fig. 14.9). On cold days residual vapour emissions of a cheese factory located 80 m distant can be observed flowing through the porous basal scoria (Carracedo and Troll 2006). The cheese factory is operative and vapour emission is still active, despite the lack of continuous strong seismicity.

Low-magnitude seismicity has been continuously recorded on the island of Tenerife. Jimenez and Garcia-Fernandez (2000) carried out an experiment from December 1987 until October 1992 using a vertical-component seismic station from the network deployed by the Estación

Volcanológica de Canarias (CSIC) at the foot of Teide volcano (3 km from the summit). Routine analysis of this station's analog records during that period shows the occurrence of microseismic activity (only recorded at this station, and thus impossible to locate), that these authors correlated with intense precipitation periods and fluid circulation through fractures. Tarraga et al. (2006) and Carniel et al. (2008) interpret the seismic noise recorded in the island of Tenerife either as anthropogenic, or due to tectonic events.

In any case, the existence of anthropogenic and meteorological noise should be considered to allow better identification and discrimination of local seismicity associated with volcanic activity, to improve eruptive surveillance and prevent false alarms.

14.3 Volcanic Hazards in the TVC

A significant difference between Hawaiian and Canarian volcanism is the abundant felsic and potentially more hazardous eruptions in the latter (Clague 1987; Ablay et al. 1998; Rodríguez-Badiola et al. 2006; and Chaps. 9 and 10 in this volume). As illustrated in Fig. 1.16, the magmatic series of the TVC includes evolved rocks (phonolites, trachytes) and, therefore, eruptive mechanisms may be explosive, as described in Chap. 12.

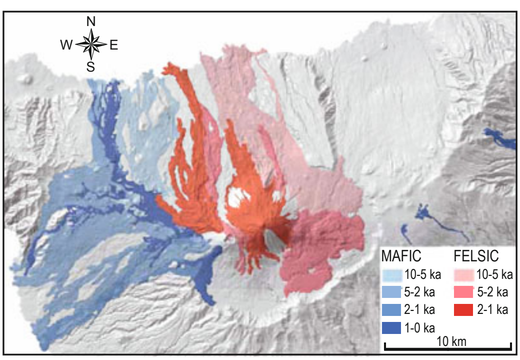

Fig. 14.10 Holocene distribution of *mafic* and *felsic* eruptions in the Teide Volcanic Complex (from Carracedo et al. 2007)

The interaction of the NW and NE rift zones of Tenerife in the construction of the post-Las Cañadas Caldera central Teide-Pico Viejo complex explains the coeval but distinctly separate mafic and felsic eruptions. The former relate to the rifts, particularly the NW rift zone, whereas the felsic volcanism is only associated to the central volcanoes, implying that explosive hazards are geographically confined (Fig. 14.10).

14.4 Lava Flow Hazards

Eruptive activity is relatively intense in Tenerife, especially in the rifts, even though the island is at present in the post-erosional rejuvenation stage (Carracedo 1994; Navarro and Farrujia 1989). During the Holocene, lava flows from the TVC have resurfaced a great part of the centre of Tenerife (Fig. 14.11). However, the potential for resurfacing large areas is considerably higher in the shield-stage islands, La Palma and El Hierro. In comparison, the greatest rate of resurfacing is found in the island of Hawaii, where historical eruptions (since 1778 A.D.) have covered most of Mauna Loa and Kilauea volcanoes (Fig. 14.12). The percentage area covered by lava in the last 750 years is >65 in the rift zones of both volcanoes (Mullineaux and Peterson 1974; Mullineaux et al. 1987).

The age and the main petrological and morphological features (Table 14.1), as well as the evaluation of cumulative volumes of lava flows (Fig. 14.13) of the recent volcanism are essential parameters to assess the eruptive hazards posed by the TVC. An important tool is a normalised lava flow resurfacing map that counts the number of lava flows on each given area of the volcano.

With this basic information, hazards related to lava flows can be summarised by answering some of the following questions: Which areas are most affected? What distance can lava flows travel considering their composition and viscosity? What volumes can be emitted and over what duration?

Calculations of the percentage area covered by lava flows in the TVC during the last 12 ky (Fig. 14.14) show that the NW rift-zone has

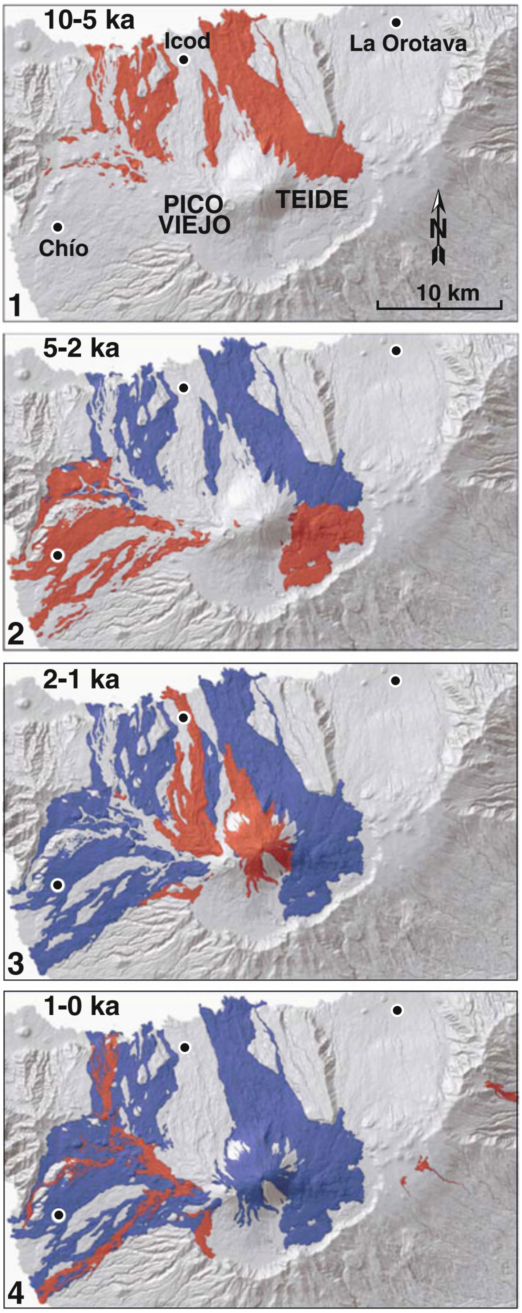

Fig. 14.11 Cumulative resurfacing of the Teide Volcanic Complex by lava flows in the last 10 ky (from Carracedo et al. 2007)

always been the most active with 95 % of its areal extent being renewed (covered) by lava flows. Despite minor historical eruptions (1704–1705), the NE rift-zone appears to be declining (resurfaced area <10 %). Almost 5.5 km^3 of lava flows were emplaced during the last 14 ky in Tenerife (Table 14.1), 1 km^3 from the NW rift-zone and 4.4 km^3 from the central edifices (Teide and peripheral domes). The thick felsic (phonolitic) flows from peripheral domes have greater volumes (e.g. 1.4 km^3 for Roques Blancos), whereas the basic lava flows of the NW rift-zone have volumes rarely exceeding 0.1 km^3 (e.g. 110×10^6 m^3 for Montañas de Chío).

Both felsic and basic flows are able to reach the coast, thus covering distances of 10–15 km. Lava flows in Hawaii reach exceptional lengths (>50 km) and eruption durations are sometimes very long (e.g., Mauna Ulu, Pu'u 'O'o), not only because of the greater rates and volumes and frequent formation of lava channels and tubes, but also because of the longer distances of eruptive centres (generally at the rift zones) from the coast in Hawaii. In both islands, the majority of lava flows reach the sea (see Figs. 14.11 and 14.12). Therefore, for hazard assessment, lava flow length is not a distinctive parameter, and hazard maps must consider that lava will flow over the "maximum" distance, i.e., all the way to the coast.

The cumulative volumes of basic and felsic lavas emitted during the last 15 ky in Tenerife is controlled by two pulses of felsic volcanism (~5000 BP and 1800–1200 BP), whereas the production of basic volcanism appears continuous through time at a rate of ca. 66×10^6 m^3/ky).

The eruptive rates of the observed historic eruptions along the NW rift-zone of Tenerife were determined using the total erupted volumes and the duration of the eruptions (Table 14.1). The high eruption rate of the 1706 Garachico eruption (71 m^3/s) is of the order of the 1669, 1980 and 1989 eruptions of Etna, and the 1977–1984 eruptions of Krafla (Harris et al. 2000; Crisci et al. 2003). The other historic eruptions in Tenerife were less productive (<13 m^3/s), similar to the 1983–1987, 1991–1993, 1999 and 2001 eruptions of Etna (Calvari et al. 1994). Considering the range of eruption rates and aspect ratios, the duration of

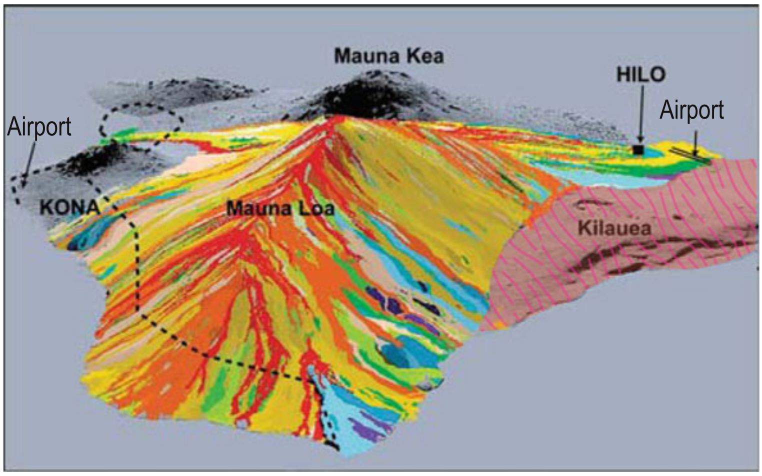

Fig. 14.12 Recent (<50 kyr) volcanism on the island of Hawaii. Note that historical eruptions (orange and red colours) have resurfaced the greater part of Kilauea Volcano and the rifts of Mauna Loa Volcano (US Geological Survey)

past eruptions of the NW rift-zone may have been typically in the order of 10 days to one month. With individual volumes greater than 500×10^6 m^3, the phonolitic eruptions of Roques Blancos, Pico Cabras, Abejera Alta and the latest eruption of Teide, the Lavas Negras, may have lasted for several months.

14.5 Hazard Maps

The main features of volcanism in the TVC served to compile a map of lava flow hazards and make comparisons with that of the island of Hawaii (Fig. 14.15). These maps need to be simple, with a limited number of zones, to capture the fundamental basis for models and risk mitigation efforts. The comparative analysis of lava flow hazard maps of both islands yields interesting conclusions. In both cases the areas with greater hazards are the rift zones. In Tenerife, significant hazards are also associated with the main central complex (Fig. 14.15a), whereas in Hawaii this is related to the main shield volcanoes, particularly the younger Mauna Loa and Kilauea (Fig. 14.15b).

In Tenerife, the greater part of the economic activity and the main settlements are concentrated in areas of old volcanism (the Mio-Pliocene shield volcanoes and the eastern part of the NE rift zone). This "spontaneous" land use is the result of agriculture having been the main economic resource of the island, with the sensible tradition of avoiding areas of recent volcanism, especially the rugged surfaces of 'a'ā flows, which are locally referred to as "*malpaíses*" (badlands).

Historically, therefore, the capital and the main populated areas, the airport, the harbour and the touristic areas were initially located in the NE of the island, on the eastern edge of the NE rift and the Orotava and Güímar Valleys.

Since 1960, massive new tourist developments have been established on the south flank of the island. This coincides with the advantage provided by the rim of the Las Cañadas Caldera, efficiently acting as an obstacle to the southward flow of lavas and small pyroclastic eruptions from the TVC, creating a topographically protected area (Fig. 14.14a).

In contrast, the economy and population of Hawaii is less protected from lava flow hazards. The capital, airports, harbours and the main tourist resorts have been built on the flanks of the very active Mauna Loa and Kilauea volcanoes (Fig. 14.15b).

In recent years there has been a growing tendency to "exaggerate" the geological hazards (seismic and eruptive) in the Canary Islands. Recent examples of these overstatements are the 2004 "reawakening of Canary Island's Teide volcano" (García et al. 2006), and the alleged potential collapse and tsunami at La Palma (Ward and Day 2001).

Table 14.1 Age and main characteristics of mafic and felsic eruptions of the Teide Volcanic Complex

Eruption	Age conventional	Age calibrated	Petrology	Location	Duration days	Area km^2	Δ alt. km	Runout length km	Aspect ratio	Slope angle degrees	Thickness (max) m	Thickness (mean) m	Volume 10^6 m^3	Eruption rate m^3/s
Chinyero	1909		Basanite	NW rift	10	2.7	0.46	4.6	2.2	5.7	6	4	10.8	12.5
Chahorra (Narices)	1798		Phonotephrite	Central	92	4.9	0.66	5.2	2.4	7.3	10	6	29.4	3.7
Garachico	1706		Basanite	NW rift	9	10.8	1.27	6.6	1.3	11.1	10	5	54.0	69.4
Arafo	1705		Basanite	NE rift	25	4.5	1.42	9.5	2.5	8.6	10	6	27.0	12.5
Siete Fuentes	1704–1705		Basanite	NE rift	6	0.3	0.16	1.7	6.5	5.4	6	4	1.2	2.3
Fasnia	1704–1705		Basanite	NE rift	21	0.9	1.12	5.5	6.5	11.8	10	7	6.3	3.5
Boca Cangrejo	350 ± 60 BP	1430–1660 AD	Tephrite	NW rift		5.8	1.34	11	1.8	7.0	7	5	29.0	
Reventada	990 ± 70 BP	900–1210 AD	Phonotephrite to phonolite	NW rift		21.8	0.93	9.9	1.1	5.4	8	6	130.8	
Volcán Negro			Tephrite to phonotephrite	NW rift		0.5	0.10	0.8	5.0	7.2		4	2.0	
Cuevas Negras			Phonoteph to phonolite	NW rift		10.8	2.45	17	1.3	8.3		5	54.0	
Teide (Lavas Negras)	1240 ± 120 BP	660–940 AD	Phonolite	Central		32.8	2.66	9.0	3.1	17.2	35	20	656.0	
Roques Blancos	1790 ± 120 BP	85–387 AD	Phonolite	Central		27.8	2.70	14.7	8.4	10.6	160	50	1390.0	
Volcán Los Hornitos	1930 ± 80 BP	39 BC–209 AD	Basanite	Central		3.2	1.05	7.7	2.5	7.8	6	5	16.0	
El Boquerón	2420 ± 140 BP	764–3942 BC	Phonolite	Central		2.1	1.24	3.5	18.3	20.8	60	30	63.0	
Los Gemelos			Phonolite	Central		0.2	0.18	0.7	39.6	14.9		20	4.0	
Mña. Mancha Ruana			Phonolite	Central		1.1	0.95	3.5	8.4	15.7		10	11	
Montañas Negras			Basanite	NW rift		4.8	1.30	7.3	2.0	10.3		5	24.0	
Volcán El Ciego	2660 ± 40 BP	900–790 BC	Phonotephrite	NW rift			1.05	11.3		5.3				
El Espárrago			Basanite	NW rift		9.2	1.36	9.7	1.5	8.1		5	46.0	
Montaña Cascajo			Basanite	NW rift		18.4	1.65	11.6	1.0	8.2		5	92.0	
Montaña Samara			Tephrite	NW rift		1.8	0.28	2.5	4.0	6.4		6	11.0	
Montaña la Botija			Basanite	NW rift		10.5	2.10	13.8	1.4	8.8		5	52.5	
Mñas. de Chío	3620 ± 140 BP	2197–1772 BC	Benmoreite	NW rift		22.0	2.06	15.6	0.9	7.6	6	5	110.0	
Montaña Bilma			Basanite	NW rift		5.8	1.20	9.6	1.8	7.2		5	29.0	

(continued)

Table 14.1 (continued)

Eruption	Age conventional	Age calibrated	Petrology	Location	Duration days	Area km^2	Δ alt. km	Runout length km	Aspect ratio	Slope	angledegrees	Thickness (max) m		
Thickness (mean) m	Volume 10^6 m^3	Eruption rate m^3/s	Montañetas Negras								Basanite	NW rift		0.2
0.06	0.8	5.9	4.3		3	0.6								
Mña. Cruz de Tea			Basanite	NW rift		3.5	0.46	5.2	2.4	5.1		5	18.0	
Mña. Majúa			Phonolite	Central		3.1	0.10	2.4	12.6	2.4		25	77.5	
Montaña Cruz Caldera			Trachyphonolite	Central		5.0	0.10	3.2	11.9	1.8		30	150.0	
Mña. Juan Evora			Basanite	Central		0.5	0.08		5.0	4.6		4	2.1	
Mña. Abejera Baja	4790 ± 140 BP		Phonolite	Central		3.4	1.06	5.6	19.2	10.9	100	40	136.0	
Mña. Abejera Alta	5170 ± 110 BP		Phonolite	Central		16.5	2.43	14.1	6.5	9.9		30	495.0	
Cueva del Ratón	5370 ± 50 BP		Phonotephrite	NW rift		15.0	1.60	9.9	1.1	9.3		5	75.0	
Pico Cabras			Phonolite	Central		15.5	2.30	12.5	11.3	10.6	110	50	775.0	
Mña. Los Conejos			Phonolite	Central		8.6	0.64	5.6	7.6	6.6		25	215.0	
Mña. Liferfe	7400 ± 40 BP		Phonotephrite	NW rift		9.1	1.50	9.1	1.5	9.5		5	45.5	
Mña. de Abeque			Mafic-basanite	NW rift			1.70	10.6		9.2				
Bocas de Dña María			Phonotephrite	Central		10.8	1.00	5.6	4.0	10.3		15	162.0	
Mña. Negra Tomillos	8220 ± 120 BP		Phonotephrite	Central		9.2	2.10	11.5	5.8	10.5		20	184.0	
Volcán del Portillo upper	11080 ± 160 BP		Phonotephrite	NE rift		4.7	1.40	8.3	4.1	9.7		10	47.0	
Volcán del Portillo lower	12020 ± 160 BP		Benmoreite	NE rift		5.3	0.90	6.2	3.8	8.3		10	53.0	
Mña. del Banco	12810 ± 60 BP		Basanite	NW rift		4.9	1.10	6.8	2.0	9.3		5	24.5	
Mña. Mostaza		15 ± 1 ky	Basanite	NE rift		2.9	0.05	4.2	2.1	0.7	6	4	11.6	
Volcán del Palmar		153 ± 6 ky	Basanite	NW rift		6.6	0.70	7	2.4	5.7		7	46.2	
Teno Alto		178 ± 6ky	Basanite	NW rift		4.4	0.70	4.5	3.0	8.9		7	30.8	
Mña. de los Silos		198 ± 8 ky	Basanite	NW rift		2	0.10	1.6	4.4	3.6		7	14.0	
Tierra del Trigo		261 ± 7 ky	Basanite	NW rift		3.0	0.70	3.9	3.6	10.3		7	21.0	
Mña. de Taco		706 ± 15 ky	Phonotephrite	NW rift		8.5	0.15	2	2.1	4.3		7	59.5	

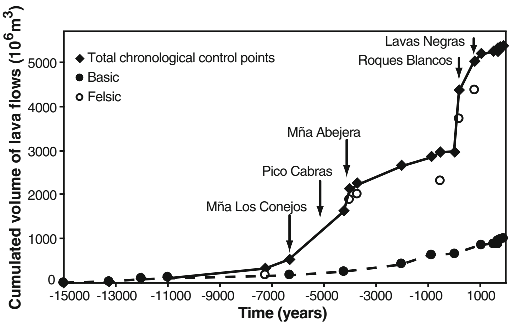

Fig. 14.13 Cumulative volumes of mafic and felsic volcanism during the last 15 ky in Tenerife. The most voluminous eruptions are indicated by arrows

Although the island of Tenerife (2,034 km^2, population 905,000 with 440 per km^2, >5 million annual visitors) is more densely populated than Hawaii (10,400 km^2, population 150.000 with 18 per km^2, 1.5 million annual visitors), the frequency and size of lava flows in Tenerife (volume, area covered, etc.) is lower than in Hawaii by about an order of magnitude. Furthermore, the historical distribution of population and economic infrastructure in Tenerife are in better "balance" with its volcanic history and eruptive hazards (Carracedo et al. 2006, 2007).

14.6 Topographic Control on Lava Flow Paths and Lava Inundation

The statement made by Klein (1982) that "Hawaiian eruptions are largely random phenomena displaying no periodicity; that is, future eruptions are relatively independent of the date of the last eruption" can be equally applied to eruptions in the Canaries, Tenerife and the TVC. In order to minimise the risk imposed by future eruptions the only means available are the detailed reconstruction of eruptive history to understand the scale of phenomena and adequate monitoring to give early warnings of imminent eruptive activity.

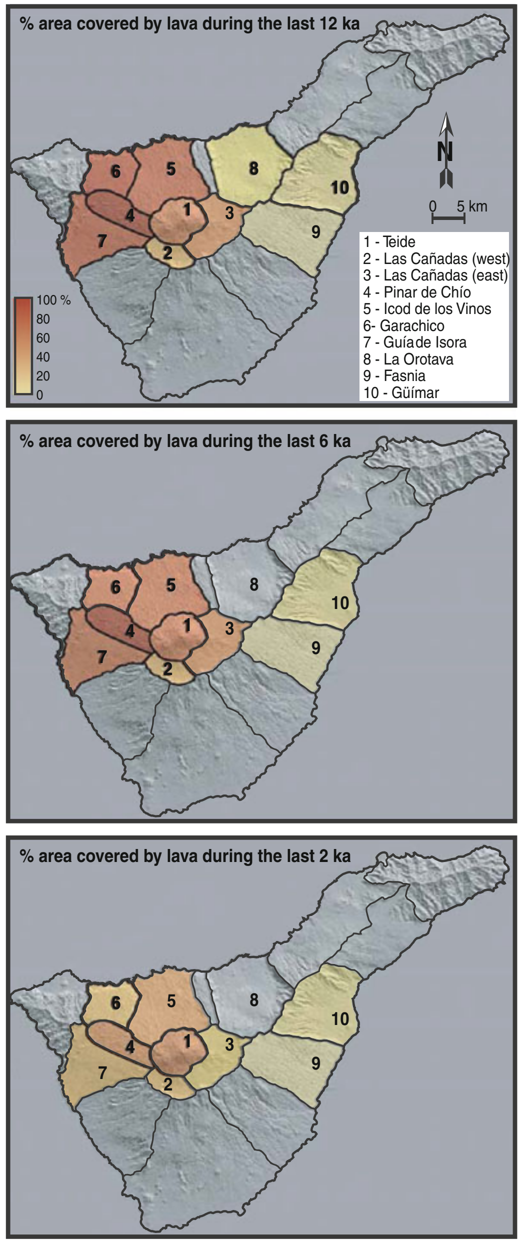

Fig. 14.14 Maps of percentage area covered by lava flows during the last 12, 6 and 2 ky, calculated after numerical mapping of the flows in a geographical information system (GIS). Main zones were defined using the topographical boundaries (hydrological basins) given and the pathways of recorded lava flows

Simulations of patterns of lava flow paths and lava inundation in different scenarios and comparison with the actual progress of well-known events (particularly historical eruptions), can provide useful information to better anticipate the behaviour of future events contrary to the

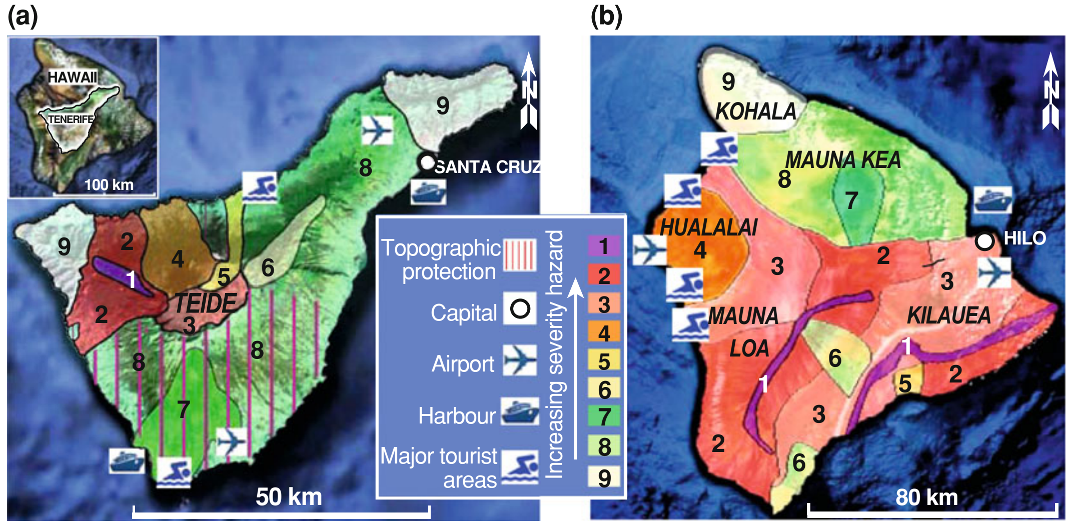

Fig. 14.15 **a** Lava flow hazard zone map of Tenerife. Hazard zones from lava flows are based chiefly on the location and frequency of Holocene eruptions (modified from Carracedo et al. 2007). **b** Current map showing volcanic hazard zones on the island of Hawaii, based on the probability of coverage by lava flows (USGS Hawaiian Volcano Observatory (HVO), http://pubs.usgs.gov/gip/hazards/maps.html)

statement by Klein (1982). Geographic Information System (GIS) and detailed geological mapping may provide valuable details to facilitate the design of strategic plans and to anticipate preventive measures.

The recently compiled GIS geological map of the TVC (Carracedo et al. 2007) provides a basis to assess the behaviour and effects of eruptions. A test of the "reliability" of these observations is to try to reconstruct historical eruptions by removing the lavas from the pre-eruption topography. Two examples are analysed here: the simulation of an eruption similar to the 1706 Garachico event (Fig. 14.16), and a simulation of the conditions required for an eruption to overflow the Las Cañadas Caldera and spill over the southern flank of the island (Fig. 14.17).

14.6.1 Inundation by a Potential Eruption Close to the 1706 Garachico Event

Figure 14.16 shows the location of a potential eruption that reproduces a situation similar to the 1706 event. Slope control will result in the partial inundation of the town of Garachico with an assumed lava thickness of 5 m reaching Garachico. Using a theoretical 10 m of lava to reach Garachico, most of the town will be flooded and the new harbour threatened. The harbour will be inundated and even buried if the eruption was to continue to the equivalent of 15–20 m thickness.

14.6.2 Overflow of the Las Cañadas Caldera

The topographic barrier formed by the wall of the Las Cañadas Caldera prevents lava flows from Teide and Pico Viejo volcanoes reaching the southern flank of the island. However, this topographic protection is incomplete because of several "*portillos*" (breaches), formed by backward erosion of the main *barrancos* (Fig. 14.17a). Some of these gaps, under 10 m high, can theoretically be overflown by volcanic eruptions similar to the 1798 Chahorra event.

A GIS simulation (Fig. 14.17b) shows that a 20 m thick layer of lava is required to allow the lava to enter Barranco de Erques directed towards the southern coast. However, even with

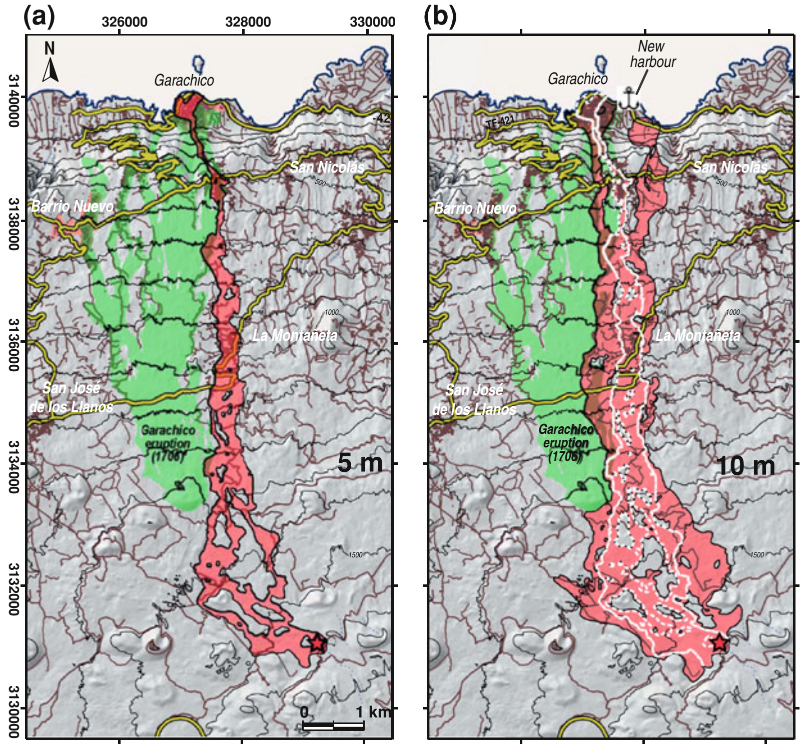

Fig. 14.16 GIS simulation of a potential eruption that could inundate the town of *Garachico*, to a similar exent as the 1706 event (**a**), and inundating it entirely (**b**)

larger volumes of lava, the hazard for the populated areas at the coast will be rather limited because the flow will still be confined inside the deep *barranco* and because the coast directly at the mouth of this *barranco* is uninhabited.

Eruptions with vents located close to the Chahorra volcano could take advantage of a similar gap at the head of Barranco del Fraile (Fig. 14.17a) to overflow the caldera rim and reach the coast. This scenario implies a more significant risk, since the village of Playa San Juan (5,000 inhabitants) is located at the mouth of this *barranco*, the foreseeable path of the lava flow. However, significant eruptive volumes would be required for this scenario.

14.7 Hazards Related to Felsic Volcanism in the TVC

Relatively frequent felsic explosive eruptions have taken place in the TVC during the Holocene (but none in the last thousand years), all of them located inside the Las Cañadas Caldera (see Fig. 14.10).

Contrasting the very explosive volcanism of the Las Cañadas Volcano (LCV) prior to the 200 ky Icod lateral collapse, the post-collapse nested construction of the TVC has overall been characterised by low-explosivity magmas. As described in Chaps. 10 and 12, despite the significant volume of felsic (phonolitic) volcanics (Ablay et al. 1998; Rodríguez-Badiola et al. 2006), there is no evidence in this latest volcanic cycle of the abundant plinian episodes of the pre-TVC activity of the Las Cañadas Volcano (>150 km^3 according to Edgar et al. 2007).

Low-explosivity eruptions, mainly mafic and felsic strombolian, characterise the TVC volcanism. However, there are some interesting examples of more violent episodes: the subplinian Montaña Blanca eruption (Ablay et al. 1995), and hydromagmatic eruptions of Pico Viejo and Calvas del Teide (Pérez Torrado et al. 2004; Pérez Torrado et al. 2006; del Potro et al. 2009). Garcia et al. (2011) reported the presence

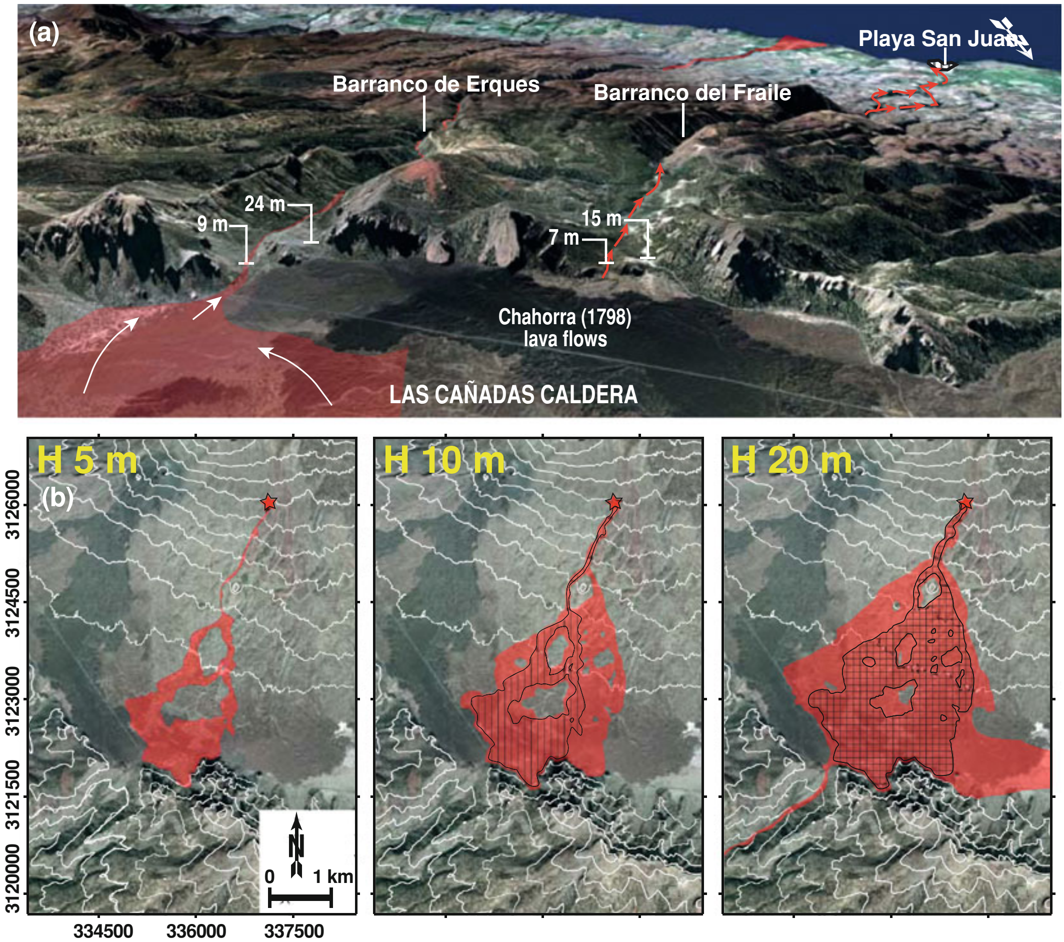

Fig. 14.17 **a** Oblique view from the NW of the southern wall of the Las Cañadas Caldera showing the breaches (portillos) carved by the backward erosion of the main *barrancos*. These gaps would be the pathways for a potential overflow of lava for future eruptions inside the Las Cañadas Caldera, threatening villages on the southern coast. **b** GIS simulation shows that for overflow to commence a 20 m thick lava flow is needed

of density current deposits, including ignimbrites and block and ash deposits, in the Holocene eruptive history of the Teide–Pico Viejo stratovolcanoes, probably associated with the gravitational collapse of domes or the front of phonolitic flows.

The Montaña Blanca eruption exemplifies the main explosive event in the TVC and the hazards that this type of volcanism presents. From this point of view, the most interesting episode of this complex (see Fig. 8.33) is the Montaña Blanca pumice fallout deposit, a single well-sorted massive bed extending northeast towards El Portillo in an elongated area reflecting southwest dominant winds during the pumice eruption (Ablay et al. 1995).

Ablay et al. (1995) carried out a detailed study of the pumice bed from sixty pits up to 3.5 m depth, constructing isopach and isopleth maps (Fig. 14.18). According to the criteria of Walker (1973) and Pyle (1989), these authors classify the deposit as subplinian, since the 0.01 T_{max} isopach greatly exceeds 40 km^2, and is probably of the order of 100–200 km^2.

According to Ablay et al. (1995), a repetition of this eruption would devastate downwind areas

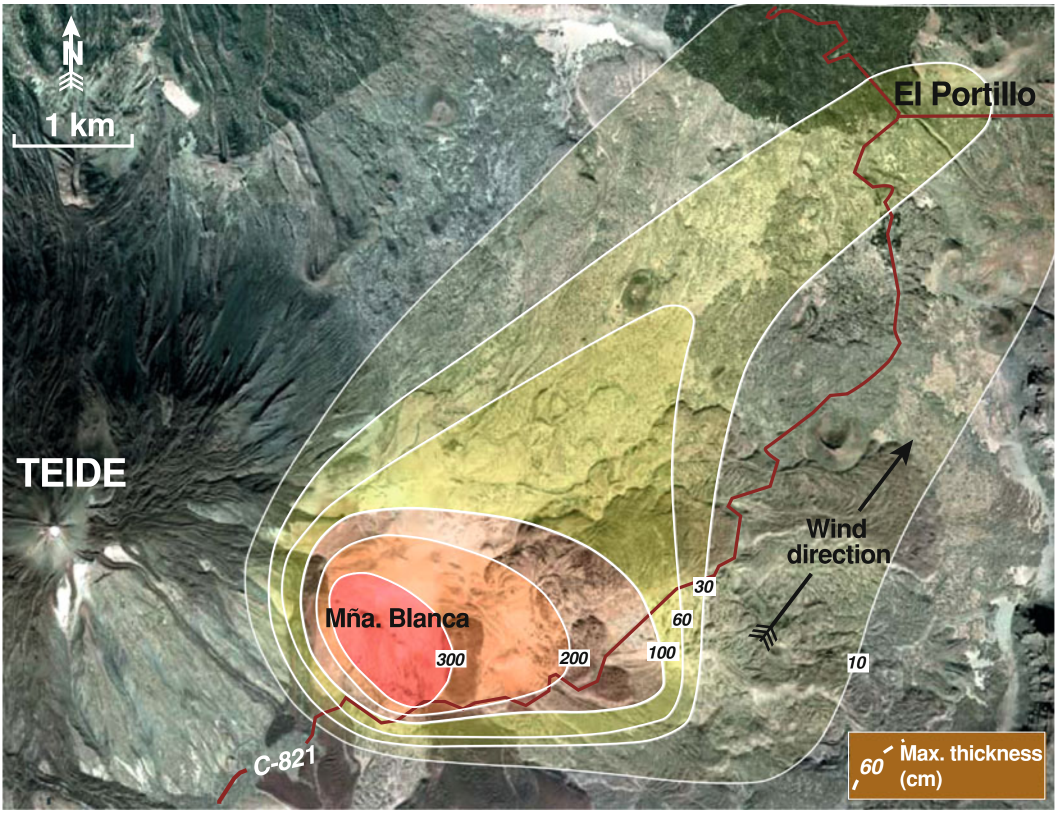

Fig. 14.18 Isopach map of the Montaña Blanca pumice deposit (contours in centimetres). Modified from Ablay et al. (1995). These authors claim that a repeat of this eruption would devastate downwind areas between the centre of Tenerife and the north-east coast within 20 km of the vent. However, the most significant adverse effects would be confined within the Las Cañadas Caldera, although some disruption may be caused at greater distances, mainly by pumice and fluorine dispersal, depending on wind direction and strength

between the centre of Tenerife and the north-east coast within about 20 km of the vent, and perhaps affect air traffic and even have widespread repercussions. However, the main impact of such an event would probably be confined inside the Las Cañadas Caldera and would depend on the wind direction and strength during the eruption.

Eruptions triggered by magma mixing have been relatively frequent in the TVC. This mechanism commonly produces very explosive events, but those recognised (e.g., Montaña Reventada) are consistently of relative low explosivity (Rodríguez-Badiola et al. 2006; Wiesmaier et al. 2011) and are restricted to the rift zone-central complex boundary (see Chap. 7).

14.8 Ground Deformation Hazards

Ground deformation is a main source of hazard in the Hawaiian Islands but is less significant in the Canaries. In Hawaii, vertical and horizontal motions of as much as 3–5 m occur in association with large earthquakes and the shallow migration and eruption of magma; slower motions of as much as 20 cm/yr are almost continuous, particularly near Kilauea's summit (Kauahikaua et al. 1994).

Ground deformation in Tenerife is very limited in comparison and related hazards appear negligible. Deformation is restricted to long-term gravitational sinking of the dense core of

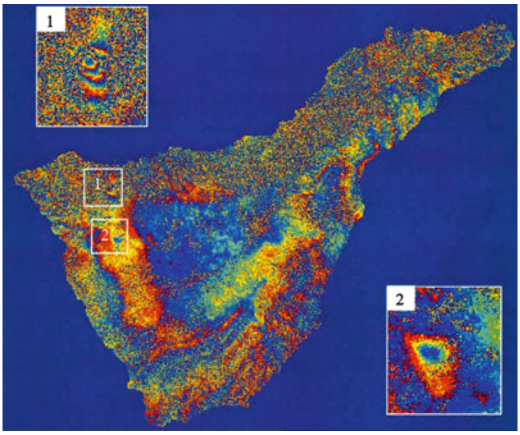

Fig. 14.19 Differential interferogram of Tenerife (1993–2000) showing two areas of subsidence interpreted to result from water table changes due to intensive extractions (from Fernandez et al. 2005)

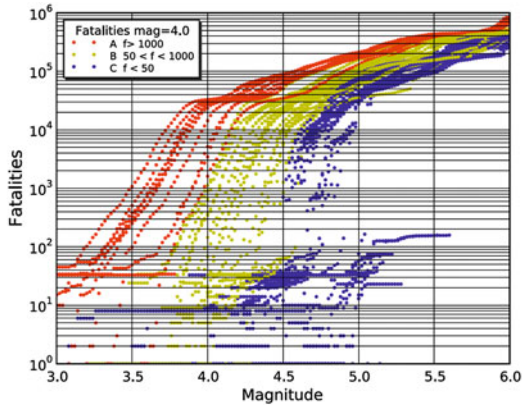

Fig. 14.20 Number of potential fatalities versus the magnitude of an eruption. The model shows that the major impacts (*red curves*) are produced by the Teide summit and north flank vents. According to this model "the emergency plan for Tenerife must provide for the evacuation of more than 100,000 people" (from Marrero et al. 2012)

the island. More localised deformation in the west of the island (1 and 2 in Fig. 14.19) has been associated with water table changes due to intensive extractions (Fernandez et al. 2005), and present a very isolated hazard only. These areas are well known and hence represent little danger to the population of the island.

14.9 The Present State of the TVC Plumbing System

In contrast to the relatively well-known volcanic plumbing system of Hawaiian volcanoes (e.g., Kilauea Volcano; Ryan 1987), very little information is available on the subcaldera magma reservoir of the TVC. Attempts to define the main features of the TVC plumbing system have led to contradictory interpretations. Albert-Beltran et al. (1986, 1990) and Diez Gil and Albert (1989) identified a central post-caldera magmatic chamber located at sea level, relatively small (~4 km in diameter) and at ~400 °C. Valentin et al. (1989) based their characterisation of this magmatic chamber on gas isotope geochemistry, concluding that the source of gases from Teide proceed from a 400 °C post-caldera magmatic chamber approximately at sea level depth underneath Teide.

In contrast, Ortiz et al. (1994) postulated a deeper magmatic chamber, almost completely solidified, sustaining that the Teide volcano is practically extinct: "The highest temperature of the 6 km deep Teide magmatic chamber is 400 °C, the final stage of cooling from the geological point of view". Araña et al. (1989) argued that this magmatic chamber is in a quasi-terminal state, probably formed by independent small magma bodies with temperatures as low as 250 °C. These authors claim that "the scientific thermodynamic models show that the present cycle of activity of Teide is in a terminal stage and, consequently, the explosivity of expected eruptions of Teide is similar to that of any other eruption in the Canaries". Araña et al. (1989) conclude that "We can assure that an eruption of Teide is virtually impossible, that it is an exhausted system". Ortiz et al. (1994 insisted in stating that "The work carried out by using petrology, fluid geochemistry, geophysics, etc., denotes the absence of magmatic chambers with the capability to produce explosive eruptions in the Teide central volcano. The hazards are magnified by the media and 'unscrupulous' research groups believing that overstating the risk will bring funds and promotion".

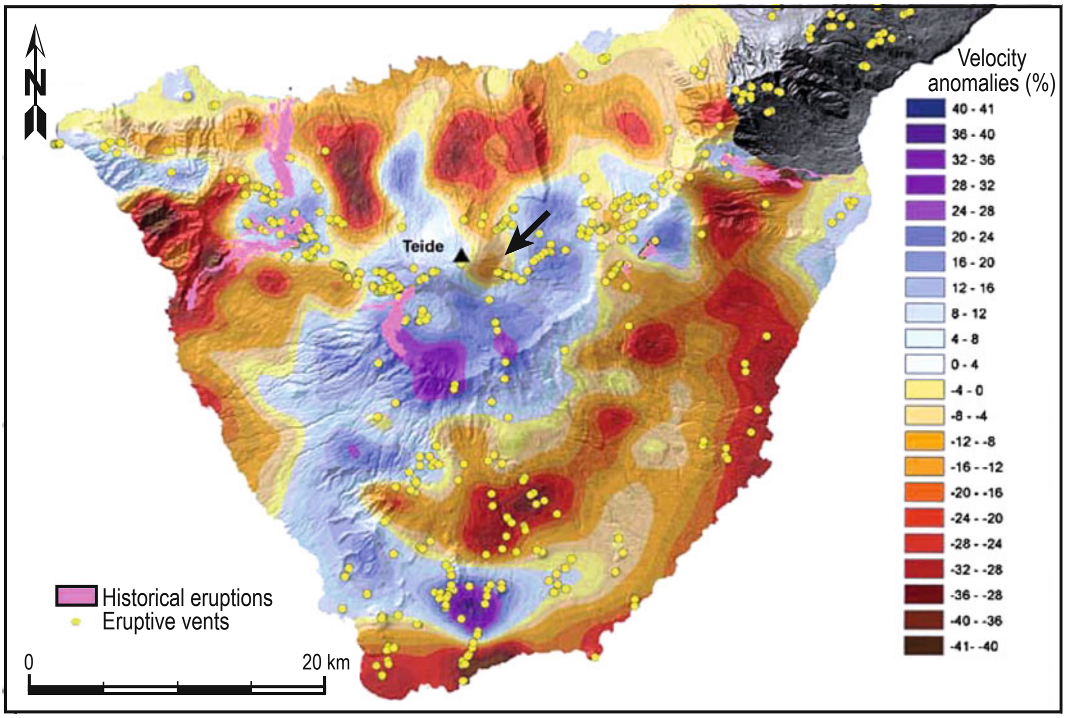

Fig. 14.21 Horizontal section at sea level of the seismic tomography model of Tenerife (from García Yeguas 2010)

This general notion changed with the increase in the number of low-magnitude events in May 2004 that was considered by some authors as unusual and pointed to a reawakening of Canary Islands' Teide Volcano (García et al. 2006; Gottsmann et al. 2006; Martí et al. 2009; Dominguez-Cerdeña et al. 2011). In a recent article, Marrero et al. (2012) assessed the impact that a possible eruption from the Tenerife Central Volcanic Complex would have on the population at risk, and urged the need to foresee "evacuating more than 100,000 persons in the case of an eruption warning in the Central Volcanic Complex of Tenerife Island. This situation would be similar to that of Vesuvius" (Fig. 14.20).

This assumption is at odds with the results obtained in a tomography study (performed using airgun shots) of Tenerife (García Yeguas 2010). In this model, the island of Tenerife shows a central nucleus of high velocities encircled by low velocity areas (Fig. 14.21). High velocity of seismic waves is commonly associated in oceanic islands to dyke swarms (rifts), plutonic bodies derived from the cooling of shallow magma reservoirs or dense cores of cumulates. The low velocity areas extending on the flanks of the island are probably associated with fractured and porous formations.

The only area of low velocity in Teide appears under Montaña Blanca in the section of the model corresponding to sea level and above (Fig. 14.22), but not in a deeper section (5,500 m b.s.l.). This has been interpreted by García Yeguas (2010) as associated with zones of hydrothermal alteration or with a shallow magma intrusion, although the latter interpretation is uncertain as only P-waves were used in this study.

14.10 The Present Risk Mitigation Challenge

Naturally, Teide poses hazards and we have now established that lava flow-, pyroclastic flow- and air-fall hazards are present at Teide (see also Martí et al., 2012). Of these, air-fall events

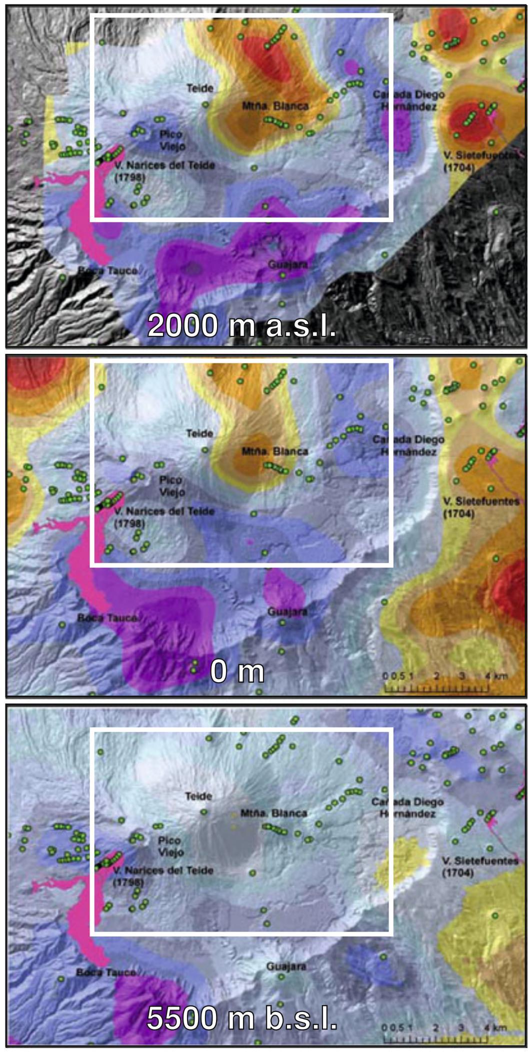

Fig. 14.22 Sections at different depths of the seismic tomography model of Tenerife. An area of low velocity appears at shallow depths in the zone of Montaña Blanca and the Abejeras-Pico Cabras lava domes (from García Yeguas 2010)

probably represent a more serious risk than for example lava flows, but the most serious of these is the potential threat of a large pyroclastic eruption (e.g., Martí et al., 2012; Marrero et al., 2012). Probabilities for such an event are low (e.g. with about 21 % likelihood during the next 100 years; Martí et al., 2012), but estimates of casualties of the order of 100,000 inhabitants are predicted for such a scenario if no precautions are taken (e.g. if no evacuation is performed).

On the other hand, large eruptions, for all we know, do not happen instantaneously, i.e. they do not occur without premonition. Volcanoes usually display periods of significant unrest prior to eruption (e.g., Carracedo et al. 2012a; Druitt et al. 2012) and large eruptions do signal their arrival. Teide, like all other volcanoes, is likely to signal frequently on its changing condition, allowing us to take precautions and move out of the way. This precursory behaviour should allow us to readily identify an impending eruptive scenario weeks to even months in advance, giving ample preparation and response time if thoroughly planned and managed well (cf. Carracedo et al. 2012a, b).

Given today's volcano-monitoring capabilities, future eruptions should cause limited damage if handled effectively and with foresight. It is noteworthy in this context that none of the historical eruptions in the Canaries have caused a single fatality as yet. It is up to us to learn to read Teide's messages correctly and prevent fatalities in the future as well to eventually achieve a life in symbiosis with the volcano, and move out of harm's way should Teide need to clear its throat again.

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