18 Cenozoic volcanism II: the Canary Islands

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The Canarian archipelago comprises seven main volcanic islands and several islets that form a chain extending for *c.* 500 km across the eastern Atlantic, with its eastern edge only 100 km from the NW African coast (Fig. 18.1). The islands have had a very long volcanic history, with formations over 20 million years old cropping out in the eastern Canaries. Thus all stages of the volcanic evolution of oceanic islands, including the submarine stage as well as the deep structure of the volcanoes, can be readily observed. Rainfall and vegetation cover are relatively low, with the exception of the island of La Palma, favouring both geological observation and rock preservation. Furthermore, the absence of surface water has promoted groundwater mining by means of up to 3000 km of subhorizontal tunnels (locally known as 'galerías'). These galerías are especially numerous in Tenerife, La Palma and El Hierro, and allow the direct observation and sampling of the deep structure of the island volcanoes without requiring expensive and indirect geophysical methods (Carracedo 1994, 1996*a*,*b*).

Since the early work of famous naturalists such as Leopold von Buch, Charles Lyell, and Georg Hartung, the Canaries have been viewed as a 'special' volcanic island group and their origin has been closely related to African continental tectonics (Fúster *et al.* 1968*a*,*b*,*c*,*d*; McFarlane & Ridley 1969; Anguita & Hernán 1975; Grunau *et al.* 1975). However, a wealth of geological data made available in recent years, especially on the western islands, has led to the conclusion that the Canaries (and the Cape Verde islands) are similar in many aspects to hotspot-induced oceanic island volcanoes, such as the Hawaiian archipelago (for recent maps and aerial photographs see info@grafcan.rcanaria.es, ign@ign.es, igme@igme.es, info@grafcan.rcanaria.es). However, although the Canarian and Hawaiian volcanoes show common constructional and structural features (rift zones, multiple gravitational

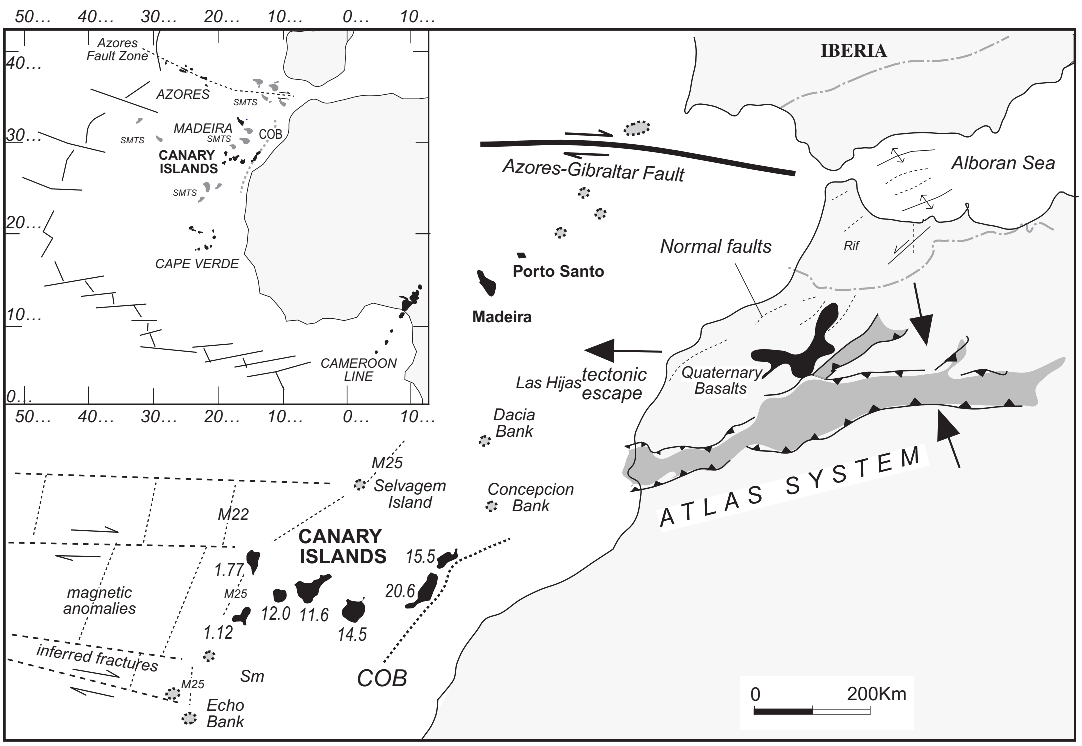

Fig. 18.1. Geographic and geodynamic framework of the NW African continental margin volcanic groups. Ages of the Canaries from Cantagrel *et al.* (1984), Ancochea *et al.* (1996), Coello *et al.* (1992), Guillou *et al.* (1996, 2001), McDougall & Schmincke 1976 (modified from Carracedo *et al.* 1998).

collapses), they also show some interesting differences in the geochemical evolution of their magmas and the amount of subsidence (Schmincke 1973, 1976, 1982; Carracedo 1984, 1999; Carracedo *et al.* 1998).

Tectonomagmatic setting, origin and evolution (JCC)

The Canary Islands developed in a geodynamic setting characterized by old (Jurassic) oceanic lithosphere lying close to a passive continental margin, and on a very slow-moving tectonic plate (the African plate). The absolute easterly motion of the African plate over fixed hotspots has been estimated (O'Connor & Duncan 1990) for the Walvis Ridge at *c.* 7° in latitude and 34° in longitude for the last 60 Ma. In the region of the Canaries these values may be as low as 2.4° and 5° (*c.* 9 mm/year), respectively, for this period.

Another potentially very important difference between the Canaries and most other oceanic island groups is that the Canaries are located adjacent to a region of intense active deformation, comprising the Atlas and Rif mountains, Alboran sea and Betic Cordillera provinces of the alpine orogenic belt. Some authors postulate the extension to the Canaries of an offshore branch of the trans-Agadir fault, associated with the Atlas system (Anguita & Hernán 1975, 1986, 2000), although there is no obvious geological or geophysical evidence for this (e.g. Dillon & Sougy 1974). Neither the Canarian archipelago as a whole, nor the individual islands and their volcanic centres and rifts follow the postulated extension of the Atlas fault to the Canaries. In fact, the islands of Fuerteventura and Lanzarote are parallel to the continental margin, whereas most of the remaining islands of the archipelago follow a general east–west trend, and the dual line of La Palma and El Hierro forms a north–south trend. Rifts in the western islands are radial and do not relate to the Atlas trend (Carracedo 1994). Rihm *et al.* (1998) demonstrate the presence of a group of apparently young seamounts (Las Hijas seamounts) located 70 km SE of El Hierro and probably destined to become the next Canarian islands. Their dyke and rift orientations are similar to those of the existing Canaries, and their location is consistent with the age-progression trend of volcanism in the Canarian archipelago and the average spacing of these islands.

The propagation of continental Atlas structures into oceanic lithosphere is likely to be mechanically unfeasible (e.g. Vink *et al.* 1984; Steckler & ten Brink 1986; ten Brink 1991). It is evident that the >150 Ma old oceanic lithosphere at the African margin in the Atlas region is considerably stronger than the continental lithosphere, precluding any fracture propagation from the Atlas towards the Canaries. Also relevant is the fact that the volcanic trend defined by the islands of Fuerteventura and Lanzarote, and an associated chain of abundant seamounts off cape Juby (Dillon & Sougy 1974), run along the continental–oceanic boundary (Figs 18.1 and 18.2). This boundary is characterized by the presence of a 10 km thick layer of sediments, and provides a zone of relative weakness and a preferential pathway for magmas (Schmincke 1982; Vink *et al.* 1984; Carracedo *et al.* 1998).

The Cape Verde islands, located 500 km off the west African continental margin, are very similar to the Canaries but exhibit a much more prominent lithospheric swell, estimated to be between 400 km and 1500 km across (Grunau *et al.* 1975) and 1500 m high at its centre (Courtney & White 1986). The apparent lack of a Canarian lithospheric swell was used by Filmer & McNutt (1988) as an argument against the presence of a hotspot in the Canaries and has also been noted by other authors (Hoernle & Schmincke 1993; Watts 1994). However, Canales & Dañobeitia (1998) analysed a number of seismic lines in the vicinity of the archipelago and demonstrated the existence of a subdued (*c.* 500 m maximum elevation) lithospheric depth anomaly around the Canary Islands. These authors proposed that this anomaly could be related to a swell that was otherwise obscured by the weight and perhaps also mechanical effects of the thick sedimentary cover along the NW African continental margin and by the weight of the volcanic rocks of the islands themselves. Swell building and magmatism for hotspots interacting with slowly moving plates are different from those seen in Hawaiian-type, fast-moving, open-ocean lithosphere, as noted by Morgan & Price (1995). Monnereau & Cazenave (1990) found that the Cape Verde, Madeira and Canary archipelagos have a much smaller island platform relief in proportion to their swell relief when compared with the Hawaiian hotspot. Satellite and surface gravity data from the Canary and Cape Verde islands (Liu 1980) show them to be the only regions of convection-generated tensional stress fields in NW Africa.

Twentieth century Atlantic intraplate seismicity (Wysession *et al.* 1995) within the African plate has been concentrated in two areas: within the Cape Verde swell and in a region of similar extent around the Canaries. The authors interpret this seismicity as being related to plume-generated magmatic activity. As suggested by Schmincke (1979), the persistence of volcanism for very long periods (>20 Ma in some islands) will require the concentration of mantle melting in a small area over a long period, an idea compatible with a mantle plume beneath the archipelago. Analysis of the seismic wave attenuation in the Canary Islands indicates zones with a high degree of attenuation and a dominance of intrinsic absorption over scattering attenuation, which points to a strong asthenosphere in the archipelago, probably hotspot-rejuvenated crust. Zones of higher values of lithospheric inelasticity are found precisely beneath the islands of Tenerife, La Palma and El Hierro (Canas *et al.* 1994, 1998). The absence of significant subsidence in the Canaries over very long time intervals (Carracedo *et al.* 1998; Carracedo 1999) may be related to slow swell evolution of the Canarian hotspot on a slow-motion plate.

Analysis of isotopic variations with distance and time in the Canaries has provided further evidence for a mantle plume origin. Hoernle *et al.* (1991) reported isotopic systematics of lavas from Gran Canaria that appear to have a plume-like composition, with high $^{238}U/^{204}Pb$. According to these authors, the plume was located to the west of Gran Canaria during the Pliocene–Recent epochs. Hoernle & Schmincke (1993) analysed major and trace elements in the island of Gran Canaria, concluding that mafic magmas were probably formed by decompression melting in an upwelling column of asthenospheric material. More recently, Hoernle *et al.* (1995) found evidence from seismic tomography and isotope geochemistry of a large region of upwelling in the upper mantle extending from the eastern Atlantic to the western Mediterranean.

The volume and distribution of the island volcanoes built above this Canarian hotspot provide interesting information about the evolution of the archipelago. The islands rest on an oceanic floor that deepens progressively westwards, reaching a depth of 4000 m in the area of La Palma and El Hierro (Fig. 18.2A). Shaded-relief images (Fig. 18.2B and C), which give

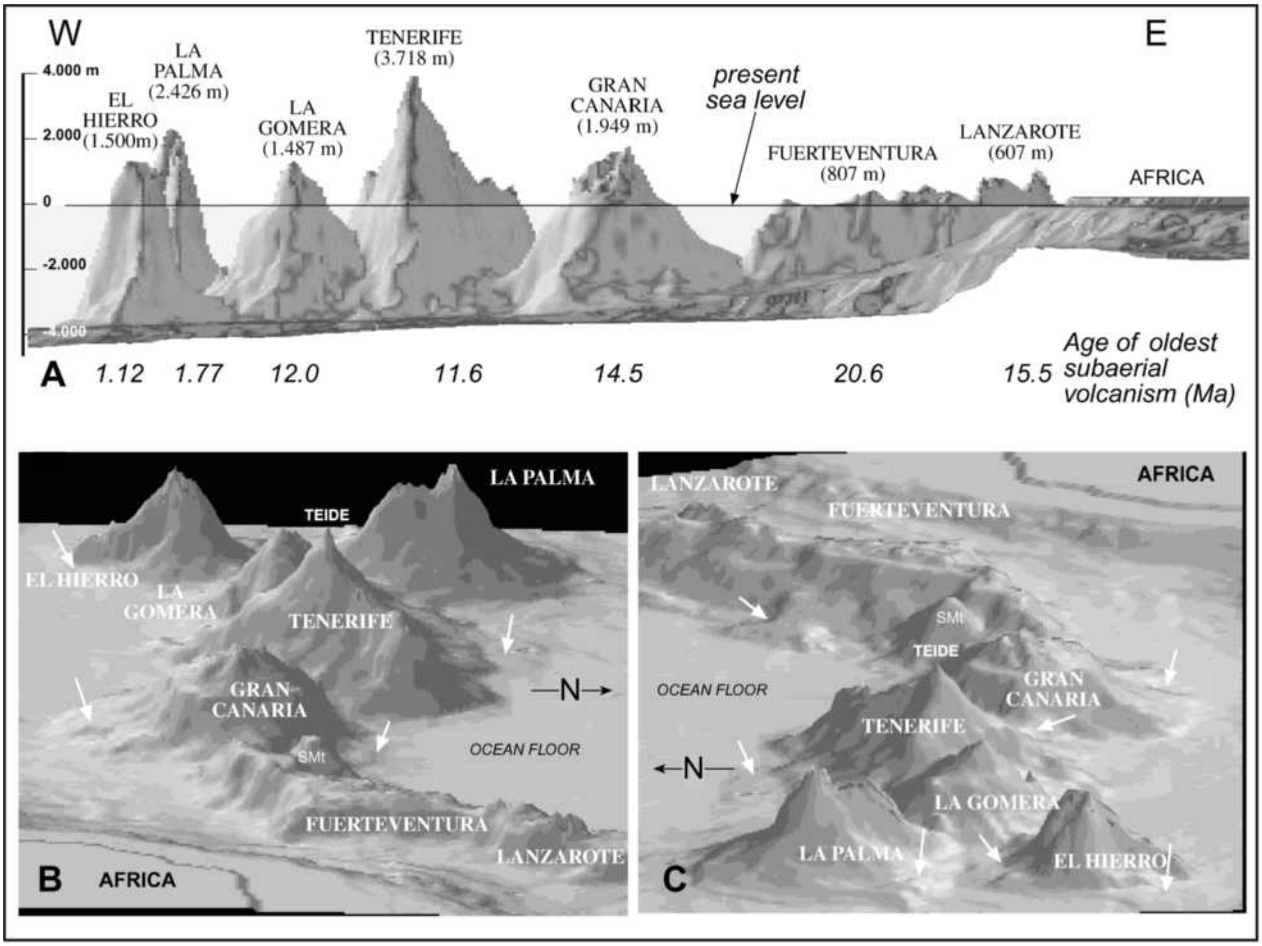

Fig. 18.2. (A) Shaded relief cross-section (east–west) of the Canaries showing the subaerial/submarine volumes of the island edifices and the corresponding oldest age of the subaerial volcanism for each island (from Carracedo 1999). **(B)** Shaded relief view of the Canaries from the east. **(C)** Shaded relief view of the Canaries from the west. Arrows indicate debris avalanche deposits from giant landslides (modified from Carracedo *et al.* 1998).

an 'empty ocean' view of the Canarian chain, clearly show that the elevation and emerged volume of the islands increase as their age decreases, with a generally westward trend. Three different groups of islands can be recognized by comparing the volume and aspect ratio of the islands with their relative ages (Fig. 18.2A): (1) Lanzarote, Fuerteventura, Gran Canaria and La Gomera are clearly older islands whose subaerial edifices have largely been mass-wasted by erosion; (2) Tenerife, the highest and most voluminous island, is probably at the peak of volcanic construction; (3) La Palma and El Hierro are both still in a very juvenile stage of growth.

Destructive processes eroding the Canaries are important in shaping island geomorphology and provide a major contribution to sea-floor sediments both surrounding the islands and on the Atlantic abyssal plains to the north. Recent deep-sea surveys and ocean drilling programmes have revealed several debris avalanche deposits around the islands (Holcomb & Searle 1991; Weaver 1991; Weaver *et al.* 1992, 1998; Watts & Masson 1995; Masson 1996; Urgelés *et al.* 1997, 1998, 1999; Teide Group 1997; Funk & Schmincke 1998). These works and related studies onshore (Carracedo 1994, 1999; Carracedo *et al.* 1998, 1999*a*,*b*; Ancochea *et al.* 1994; Guillou *et al.* 1996, 2001; Stillman 1999) have shown giant landslides to be a dominant feature of the islands' erosion.

Large-scale distribution and age progression in most oceanic island chains are well explained by the steady movement of lithospheric plates over fixed mantle plumes, yielding chains of consecutive discrete volcanoes. In this model, a new island starts to form when the bulk of the previous one has already developed, the inter-island distance being governed by lithospheric thickness and rigidity (Vogt 1974; ten Brink 1991). At the initiation of an oceanic island chain the first magma intrudes through unloaded pristine lithosphere. As the first island develops the volcanic load grows and increasing compressive flexural stresses eventually exceed the tensile stresses of the mantle plume, blocking magma pathways to the surface. This precludes the initiation of new eruptions in the vicinity of any new island volcano (ten Brink 1991; Carracedo *et al.* 1997, 1998; Hieronymus & Bercovici 1999).

An interesting feature of the Canarian island chain is the fact that the islands of La Palma and El Hierro are growing simultaneously and form a north–south trending dual line of island volcanoes, perpendicular to the general trend of the archipelago (Fig. 18.2B and C; see also Fig 18.14A and B).

Dual-line volcanoes, such as the Kea and Loa trends in the Hawaiian islands, have been associated with changes in plate motion, resulting in the location of a volcanic load off the hotspot axis. Compressive stresses related to the off-axis volcano block the formation of the next island and split the single line of volcanoes into a dual line of alternating positions of volcanoes (Hieronymus & Bercovici 1999).

In the Canarian chain, the association of a dual line of volcanoes with a change in direction of the African plate is not clear since such a change took place between the formation of Gran Canaria and La Gomera (14.5 to 12 Ma BP). A possible model to explain the splitting of the Canaries in a dual line may be related to the anomalous position of La Gomera, apparently older than Tenerife but ahead in the hotspot pathway. In fact, as pointed out above, the islands of Lanzarote, Fuerteventura, Gran Canaria and La Gomera (old Canaries) seem to form a different group from Tenerife, La Palma and El Hierro (young Canaries). Some perturbation in plate motion may have resulted in the initiation of volcanism off the previous hotspot pathway, forming the island of Tenerife NE from La Gomera, compressive stresses from which may have forced the splitting of the archipelago to form a dual line of island volcanoes (Carracedo *et al.* 1997, 1998; Carracedo 1999).

In summary, the timing of eruptive activity in the islands, their morphological and structural features, seismic signature and geochemical evolution (Schmincke 1973, 1976, 1982; Carracedo 1975, 1999; Hoernle *et al.* 1991, 1995; Hoernle & Schmincke 1993; Carracedo *et al.* 1997, 1998; Canas *et al.* 1994, 1998) clearly converge on a slow-moving hotspot model.

Fuerteventura (RC, AA)

Fuerteventura is the easternmost and second largest (1677 km^2 including the Isla de Lobos) of the Canary Islands, and the closest (100 km) to the African coast (Figs 18.1 and 18.2). Together with Lanzarote and the Conception Bank, it lies along a SSW–NNE volcanic line (East Canary Ridge) that runs parallel to the continental margin, and is built on a 3000 m deep ocean floor. Fuerteventura and Lanzarote are not geologically different islands *sensu stricto* since they are separated by only a narrow stretch of sea (La Bocaina) which is shallower than the lowest sea level at glacial maxima (Fig. 18.2).

The topography of Fuerteventura is characteristic of a post-erosional island with abundant Quaternary deposits (beach sand, dunes, scree, alluvium) and minor rejuvenation volcanism (Fig. 18.3). It is predominantly smooth, with broad valleys separated by sharp erosive interfluves (locally known as 'cuchillos' or knives). The bulk of the island is predominantly of low altitude (<500 m), except in the southern peninsula of Jandía, where the maximum elevation of the island, Pico de La Zarza, reaches 807 m. The leeward (western) coastline is low, with common shell sand beaches and dunes, in contrast to high cliffs on the windward (eastward) side of the island.

The geological history of Fuerteventura is the most complex and prolonged of the Canary Islands (Fúster *et al.* 1968*b*, 1980; Stillman *et al.* 1975; Le Bas *et al.* 1986; Coello *et al.* 1992; Ancochea *et al.* 1996; Steiner *et al.* 1998; Balogh *et al.* 1999). The island is located on thick (>11 km), strongly reworked oceanic crust, and four main lithological units are exposed (Fig. 18.3): Mesozoic oceanic crust, submarine volcanic complexes, Miocene subaerial volcanic complexes, and Pliocene–Quaternary sedimentary and volcanic rocks.

Mesozoic oceanic crust

This fragment of Mesozoic oceanic crust comprises a thick (*c.* 1600 m) sedimentary sequence that rests on tholeiitic N-MORB basalts of Early Jurassic age (Robertson & Stillman 1979*a*; Robertson & Bemouilli 1982; Steiner *et al.* 1998). The basaltic rocks form sheet flows, pillow lavas and breccias and represent the only outcropping Mesozoic oceanic basement described so far in the central Atlantic. The Mesozoic sedimentary sequence is Early Jurassic to Late Cretaceous in age and consists of terrigenous quartzitic clastic sediments, black shales, redeposited limestones, marls and chalks with chert nodules. The succession is part of a deep-sea fan deposited on the west African continental margin (Fúster *et al.* 1968*b*; Robertson & Stillman 1979*a*; Robertson & Bemouilli 1982; Steiner *et al.* 1998) and is subvertical to overturned. Cleavage–bedding relationships and the orientation of minor folds indicate the presence of a major NE facing reclined fold (Robertson & Stillman 1979*a*). This fold has been interpreted as having been generated by dextral motion along a N-S orientated transcurrent shear zone (Robertson & Stillman 1979*a*) or northward-directed thrust faulting (Gutiérrez 2000).

Submarine volcanic complexes

The Mesozoic succession is unconformably overlain by a submarine volcanic complex of uncertain age (Robertson & Stillman 1979*b*; Le Bas *et al.* 1986; Stillman 1987, 1999; Gutiérrez 2000). While some authors suggest that the earliest pillow lavas are probably Palaeocene to lower Eocene, and that the build-up of the island continued up to early Miocene times (Robertson & Stillman 1979*b*; Le Bas *et al.* 1986; Stillman 1987, 1999; Balogh *et al.* 1999), others consider that the main period of submarine construction of the island was concentrated in Oligocene times (Fúster *et al.* 1980; Cantagrel *et al.* 1993; Sagredo *et al.* 1996).

This tectonically uplifted submarine volcanic complex is dominated by pillow lavas and hyaloclastites of basaltic and trachybasaltic composition, exposed in the western part of the island, but deeply eroded and variably dipping. Near the Ajuy valley the lowest volcaniclastic sediments and volcanic breccias are inverted or steeply dipping, in general conformity with the stratigraphically underlying Mesozoic sedimentary rocks (Robertson & Stillman 1979*b*; Le Bas *et al.* 1986). In contrast, to the north near El Valle, the middle and upper parts of the submarine sequence are only gently inclined to the west or to the east (Fúster *et al.* 1968*b*; Robertson & Stillman 1979*b*; Le Bas *et al.* 1986; Gutiérrez 2000). The submarine volcanic rocks are unconformably overlain by littoral and shallow-water marine deposits (reefal bioclastic sediments, beach sandstones and conglomerates), gently inclined to the west and representing the transition from submarine to subaerial activity (Fúster *et al.* 1968*b*; Robertson & Stillman 1979*b*; Le Bas *et al.* 1986; Gutiérrez 2000).

Associated with the submarine volcanic complex is a sequence of plutonic and hypabyssal intrusions that have been subdivided into an early syenite–ultramafic series, and later syenite–carbonatite complexes. The former has been called the Tierra Mala Formation (Fm) by Le Bas *et al.* (1986), or the A1 rock group by Balogh *et al.* (1999). It comprises ultramafic and mafic rocks (alkali pyroxenites, hornblendites and amphibole gabbros) intruded by *c.* 65 Ma ($^{39}Ar/^{40}Ar$; Balogh *et al.* 1999) syenites (Fúster *et al.* 1980; Le Bas *et al.* 1986; Ahijado

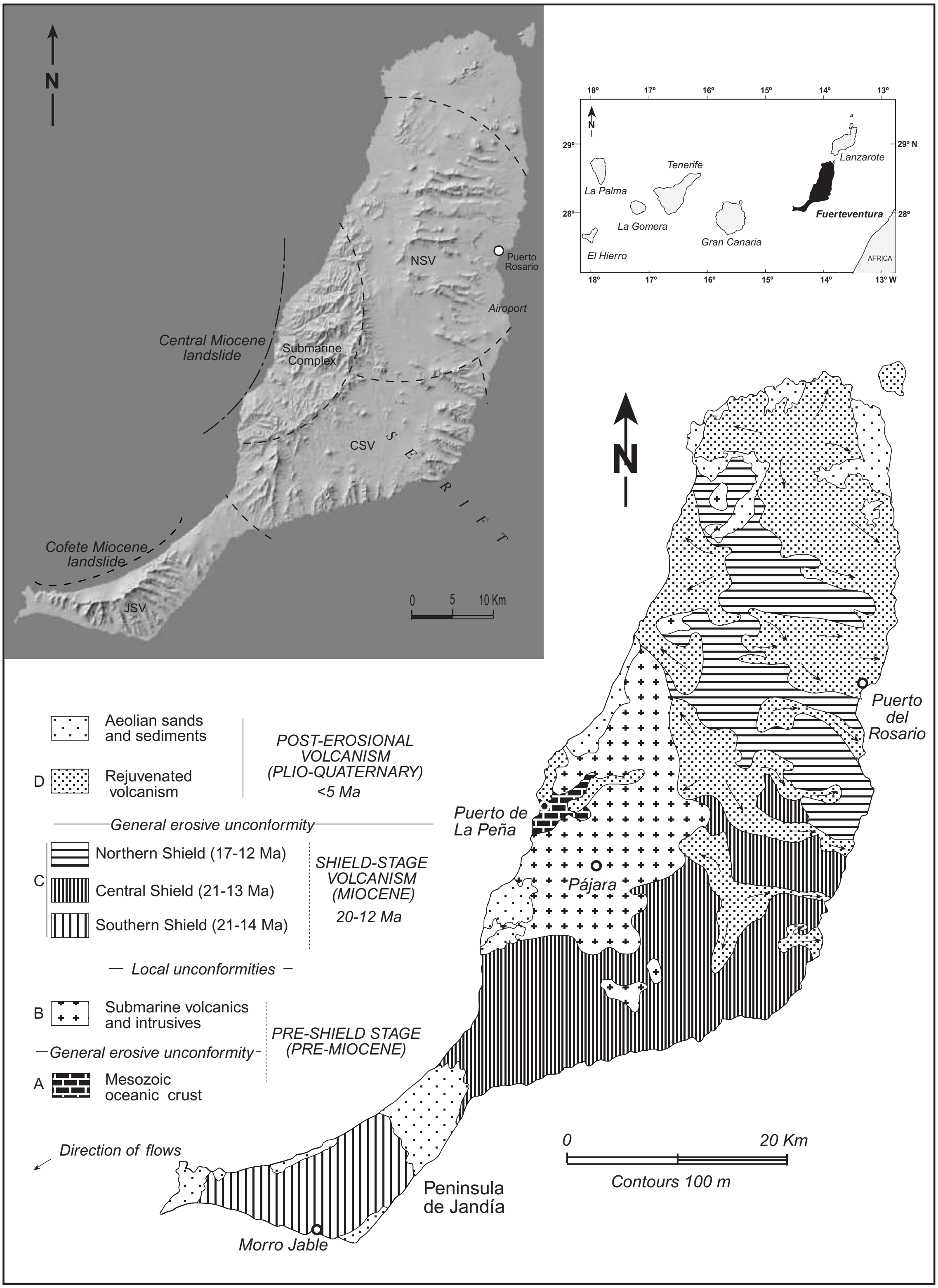

Fig. 18.3. Simplified geological map of Fuerteventura. Modified from Ancochea *et al.* (1996). Inset: shaded relief image of Fuerteventura with indication of the main geomorphological and tectonic features (image GRAFCAN). Giant landslides after Stillman (1999). Shield volcanoes: CSV, central shield volcano; SSV, southern shield volcano; NSV, northern shield volcano (after Ancochea *et al.* (1996)).

1999), and is exposed along the western coast of the island (Tostón Cotillo, Ajuy and Punta del Peñon Blanco). The later syenite–carbonatite rocks form three complexes which from north to south are: Esquinzo, Ajui-Solapa and Punta del Peñon Blanco (Fúster *et al.* 1980; Barrera *et al.* 1981; Le Bas *et al.* 1986; Ahijado 1999; Balogh *et al.* 1999). Their emplacement took place during a relatively short period of time around 25 Ma BP (Le Bas *et al.* 1986; Cantagrel *et al.* 1993; Sagredo *et al.* 1996; Ahijado 1999; Balogh *et al.* 1999). Ijolite crops out in Esquinzo and Caleta de la Cruz, and younger nephelinite dykes also appear. The emplacement of these rocks is coeval with the occurrence of several upper Oligocene–lower Miocene brittle–ductile to ductile shear zones (Casillas *et al.* 1994; Fernández *et al.* 1997*a*) arranged in a nearly orthorhombic pattern, with kinematic criteria indicating east–west horizontal extension. Shear zone activity took place after earlier deformation had affected the submarine sequence.

Miocene subaerial volcanic complexes

The early stages of subaerial growth of the island are the result of the formation of three different adjacent large basaltic volcanic complexes (Fig. 18.3 inset): the southern (SVC), central (CVC) and northern (NVC) edifices of Ancochea *et al.* (1992, 1996). They comprise mainly alkali basaltic and trachybasaltic lavas, with minor trachytic differentiates, and interbedded pyroclasts with abundant subvertical dykes. The limited amount of pyroclastics and abundant feeder dykes indicate the mostly low eruptive explosivity typical of shield volcanoes.

Each complex has its own prolonged history that may be longer than 10 Ma, during which time several periods of activity alternated with gaps characterized by major erosive episodes and giant landslides (Ancochea *et al.* 1996; Stillman 1999). The ages of these volcanic complexes (Ancochea *et al.* 1996) differ slightly. The CVC is the oldest with the bulk of construction occurring between 22 and 18 Ma BP, followed by a significant pause, and then a later smaller phase of activity from 17.5 to 13 Ma BP. In the NVC, the main activity took place between 17 and 12 Ma BP, and in the SVC volcanism occurred between 21 and 14 Ma BP. It has been estimated that the CVC volcano may have reached an elevation of over 3000 m (Javoy *et al.* 1986; Stillman 1999).

The deeply eroded remnants of these volcanic complexes crop out over much of the central E and SE part of the island and in the Jandía peninsula to the SW (Fig. 18.3). In the core of both the NVC and the CVC there are plutonic rocks (pyroxenites, gabbros and syenites) and a basaltic-trachybasaltic dyke swarm which represent the hypabyssal roots of the evolving subaerial volcanic complexes (Ancochea *et al.* 1996; Balogh *et al.* 1999). The dykes are extremely abundant, mostly orientated NNE–SSW (although some are NE–SW and others NW–SE), and seem to have involved crustal stretching of around 30 km (Fúster *et al.* 1968*b*; López Ruiz 1970; Stillman 1987; Ahijado 1999). The plutonic rocks can be subdivided into earlier gabbros and pyroxenites, and later gabbro–syenite–trachysyenite complexes. The former occur as several NNE–SSW elongated bodies (Gastesi 1969) that produce high-grade thermal metamorphic effects on the host rocks (Stillman *et al.* 1975; Hobson *et al.* 1998; Ahijado 1999), and have been dated (K-Ar) at around 20–21 Ma BP (Sagredo *et al.* 1996; Balogh *et al.* 1999). The later plutons include concentric intrusions of gabbros and syenites which form the Vega de Río y Palmas ring complex, dated at 18.4–20.8 Ma BP (K-Ar; Le Bas *et al.* 1986; Cantagrel *et al.* 1993). These intrusions were emplaced along conical fractures associated with the relaxation of the stress field that produced the dyke swarm. Other late plutonic bodies include syenites and trachysyenites of the Betancuria subaerial complex (13–14 Ma BP; Cantagrel *et al.* 1993; Muñoz & Sagredo 1996) and the Morro del Sol and Morro Negro gabbros.

Metamorphism has affected most of the rocks which form the Mesozoic oceanic crust, the submarine volcanic complex, the lower part of the subaerial volcanic complexes, and the plutonic bodies and dyke swarms related to these complexes. An intense hydrothermal metamorphism in epidote–albite greenschist facies was probably produced by the massive intrusion of dyke swarms (Fúster *et al.* 1968*b*; Stillman *et al.* 1975; Robertson & Stillman 1979*b*; Javoy *et al.* 1986). Additionally, contact metamorphism affected the host rocks of the plutons related to the subaerial volcanic complexes. In some places, near the plutonic bodies, this metamorphism reached pyroxene hornfels facies and led to the partial melting of the host rocks (Stillman *et al.* 1975; Muñoz & Sagredo 1994; Hobson *et al.* 1998).

Plio-Quaternary sedimentary and volcanic rocks

After Miocene magmatism, the subaerial volcanoes were deeply eroded, losing a large part of their original volume. Magmatic activity recommenced with the formation of Pliocene lava fields associated with basaltic volcanic vents aligned along fractures, and remaining active in prehistoric times (Cendrero 1966; Coello *et al.* 1992). Pliocene–Quaternary sedimentation produced littoral and shallow-water marine deposits overlain by aeolian deposits with intercalations of alluvial fans and palaeosols (Meco 1991).

Lanzarote (JCC, ERB)

The island of Lanzarote lies at the NE edge of the Canarian chain just 140 km from the African coast (Fig. 18.1). Elongated in a NE–SW trend, parallel to the continental margin, it is 60 km long and 20 km wide and covers 862 km^2 (905 km^2 including the small northern islets of Graciosa, Mña. Clara and Alegranza). The topography is characteristic of post-erosional islands (Fig. 18.4 inset), with deeply eroded volcanoes, broad valleys ('barrancos'), precipitous cliffs, and wide lowlands covered with aeolian sands locally called 'jables'. The highest elevation is 670 m (Peñas del Chache, in the northern part of the island) and, unlike neighbouring Fuerteventura, the island has seen volcanic eruptions in historical times.

The spectacular lava fields of the 1730–1736 eruption and the presence of high-temperature fumaroles (<600°C) were the subject of most of the early geological reports (Hausen 1959; Hernández Pacheco 1960; Bravo 1964). Marine deposits and raised beaches were studied by Meco (1977), and the first general geological study, including a geological map of the island, was published by Fúster *et al.* (1968*c*). Radiometric dating and geomagnetic stratigraphy were published by Abdel-Monem *et al.* (1971), Coello *et al.* (1992), Carracedo & Soler (1992) and Carracedo & Rodríguez Badiola (1993). More recently the historical eruptions (1730 and 1824) were studied by Carracedo & Rodríguez Badiola (1991) and Carracedo *et al.* (1992).

The geology of Lanzarote and Fuerteventura is very similar,

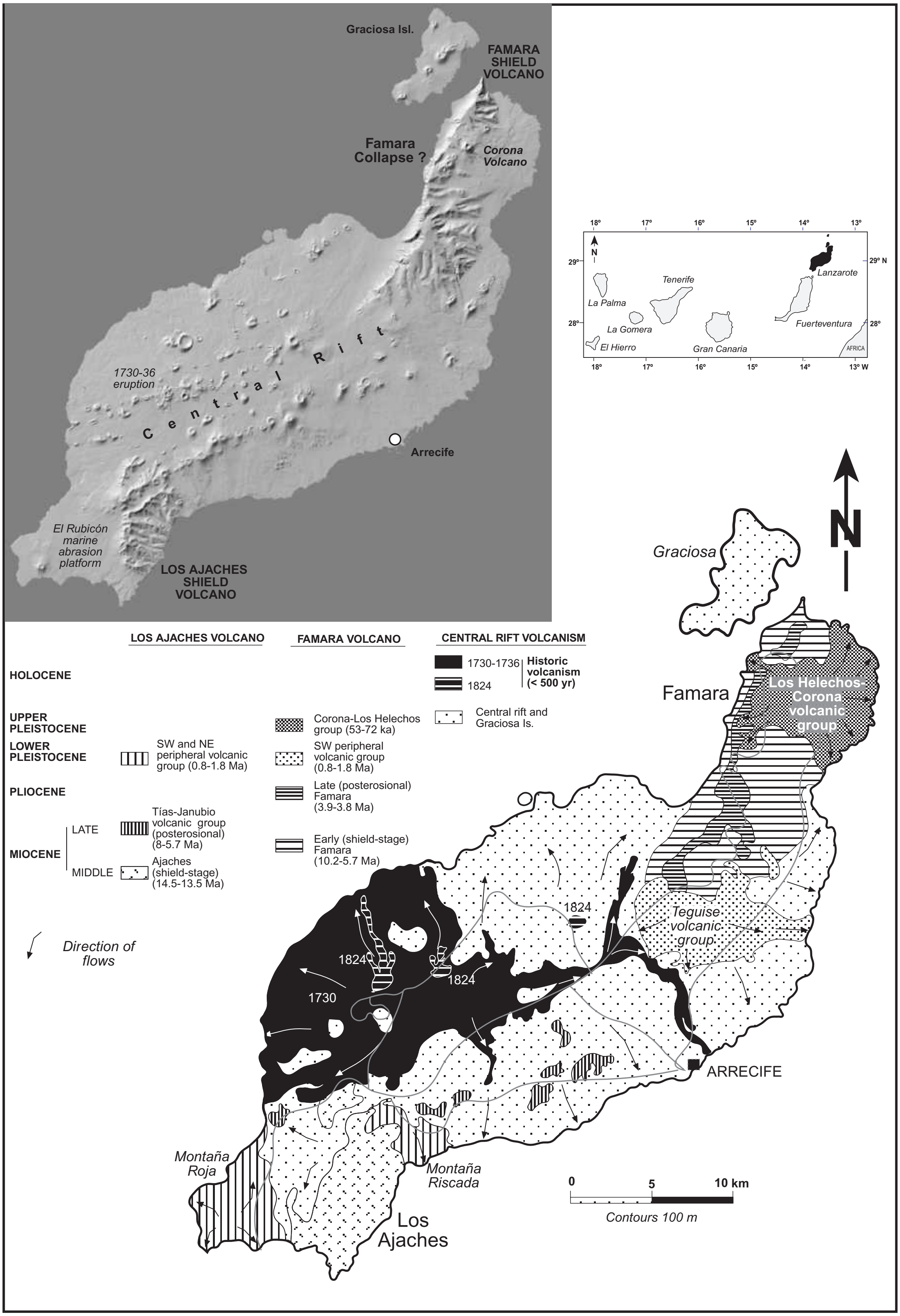

Fig. 18.4. Simplified geological map of Lanzarote. Modified from Fúster *et al.* (1968*c*). Inset: shaded relief image of Lanzarote with indication of the main geomorphological and tectonic features (image GRAFCAN).

both islands being in a very advanced post-erosional stage of development. Only the lack of significant subsidence in the Canaries (Carracedo 1999) explains why these islands are still emergent, although they have lost most of their original bulk due to catastrophic mass wasting and marine and meteoric erosion. In Lanzarote the geology is dominated by two main shield volcanoes (the Ajaches and Famara volcanoes) which developed as independent island volcanoes (Fúster *et al.* 1968*c*; Carracedo & Rodríguez Badiola 1993) and were subsequently connected by eruptive products from a central rift (Fig. 18.4). The remnants of the Ajaches volcano are preserved as basanite–alkali basalt–trachyte lavas located in the south of the island, and have been dated (K-Ar) as 19.0–6.1 Ma BP (Abdel-Monem *et al.* 1971) and 15.5–6.6 Ma BP (Coello *et al.* 1992). Comparison of the geomagnetic polarities with the established GPTS (Carracedo & Rodríguez Badiola 1993) suggests that the main shield-building stage of this volcano extends from *c.* 14.5 to 13.5 Ma BP, with volcanism from *c.* 8.0 to 5.7 Ma BP corresponding to the main post-erosional stage of the volcano, with minor later rejuvenation volcanism (Fig. 18.4).

The Famara shield volcano forms a 24 km long, NE–SW trending elongate edifice located at the northernmost part of the island (Fig. 18.4). The western half of the volcano is bound by a 600 m cliff, most probably representing the retrograded scarp of a giant gravitational collapse that mass wasted that portion of the volcano. Published radiometric (K-Ar) ages range from 10.0 to 5.3 Ma (Abdel-Monem *et al.* 1971) or 10.2 to 3.8 Ma (Coello *et al.* 1992). Geomagnetic stratigraphy suggests a main shield-building stage from 10.2 to 5.7 Ma BP and a main post-erosional stage from 3.9 to 3.8 Ma BP, with some Quaternary residual rejuvenation activity (Carracedo & Soler 1992; Carracedo & Rodríguez Badiola 1993). This trend suggests an alternation of the main eruptive phases of both the Famara and Ajaches volcanoes, similar to that proposed for the islands of La Palma and El Hierro (Carracedo *et al.* 1999*a*,*b*). The composition of the Famara volcano lavas is predominantly basanitic, with minor gradation to alkali basalts.

Recent volcanism

Quaternary volcanism is represented by peripheral eruptions around the main shields, producing the Mña. Roja and Caldera Riscada volcanoes in the Ajaches shield and the Teguise and Corona volcanic groups in the Famara shield (Fig. 18.4). The Corona lavas advanced along a wave-cut platform located 70 m below the present sea level. This eruption produced one of the largest known lava tubes in the world, 6.8 km long with sections reaching 25 m in diameter, 2 km of which are at present submerged (−80 m) following sea-level rise after the last glaciation.

Holocene eruptions are very few, probably limited to the historical 1730 and 1824 events. The 1730–1736 eruption (Fig. 18.4) is the second largest basaltic fissure eruption recorded in historical time (after the 1783 Laki eruption in Iceland). It lasted for 68 months and produced a total lava volume of *c.* 700 $m^3 \times 10^6$, compared with less than three months and 66 $m^3 \times 10^6$ for the second largest historical eruption in the Canaries. Over 30 volcanic vents were formed in five main multi-event eruptive phases, aligned along a 14 km long, N80°E-trending fissure (Carracedo *et al.* 1992). Precise reconstruction of this eruption was greatly facilitated by detailed eye witness accounts, such as the report of the parish priest of Yaiza (included in the work of Buch 1825), and particularly the official reports of the local authorities to the Royal Court of Justice, found on file in the Spanish Archivo General de Simancas (Carracedo *et al.* 1990, 1992; Carracedo & Rodríguez Badiola 1991).

An interesting feature of the 1730 activity was the eruption of transitional lavas from olivine melanephelinites through basanites and alkali basalts to tholeiites, suggesting that the lava compositions represent nearly unmodified primary melts from a heterogeneous mantle source (Carracedo *et al.* 1990; Carracedo & Rodríguez Badiola 1991, 1993). Isotopic variability found in these lavas may be partly explained as well by the melting or melt-mixing of this composite mantle source (Sigmarsson *et al.* 1998). Finally, in 1824, a short eruption from three vents aligned along a N70°E, 13 km long fissure, produced small amounts of basaltic lavas (Fig. 18.4).

Gran Canaria (FJPT, JCC)

Gran Canaria lies at the centre of the Canarian chain (Fig. 18.1) and is the third largest island (1532 km^2) of the archipelago, 45 km in diameter with a maximum elevation of 1950 m (Pico de las Nieves). It is a typical post-erosional island with important rejuvenation volcanism, and has a geomorphology that changes greatly from the NE (predominantly post-erosive volcanism) to the SW (exclusively shield-stage volcanism). A dense radial network of deep barrancos dissects the island, forming a rugged topography. The coastal landforms vary considerably, reflecting the above-mentioned differences in age, with sheer, vertical, high cliffs in the western part and coastal platforms and wide beaches and dunes in the southern and eastern regions (Fig. 18.5).

Numerous geological studies have been carried out on Gran Canaria since the eighteenth century (see Bourcart & Jeremine 1937; Hausen 1962; Fúster *et al.* 1968*d*; Schmincke 1968, 1976, 1993; and references therein), and geological maps of the island have recently been published (ITGE 1990, 1992). Rocks belonging to the seamount stage do not crop out, but oceanographic work on the volcaniclastic apron around the island was carried out during the Ocean Drilling Program (ODP) Leg 157 (Weaver *et al.* 1998). Seismic and bathymetric profiles indicate that the submarine stage formed at least 90% of the bulk volume of the island (Schmincke & Sumita 1998). Boreholes up to 300 m deep reveal graded hyaloclastite tuffs and debris flow deposits, interpreted as being derived from shallow submarine eruptions (Schmincke & Segschneider 1998). No apparent unconformity separates these submarine deposits from those of the subaerial volcanism, the latter commonly being interbedded with the former. A similarity in geochemical composition suggests a common magmatic source for both submarine and subaerial volcanic products (Schmincke & Segschneider 1998), the sole difference being in eruptive mechanisms. Once subaerial volcanism had become fully established, the onland geology of Gran Canaria records the growth of a shield volcano (evolving through shield to caldera and post-caldera stages), followed by *c.* 3 Ma of erosion, after which there was renewed magmatic activity.

Shield growth

Activity during this stage produced a complex shield volcano, 60 km in diameter, over 2000 m high, with a volume of over 1000 km^3, and occupying both the present extent of the island

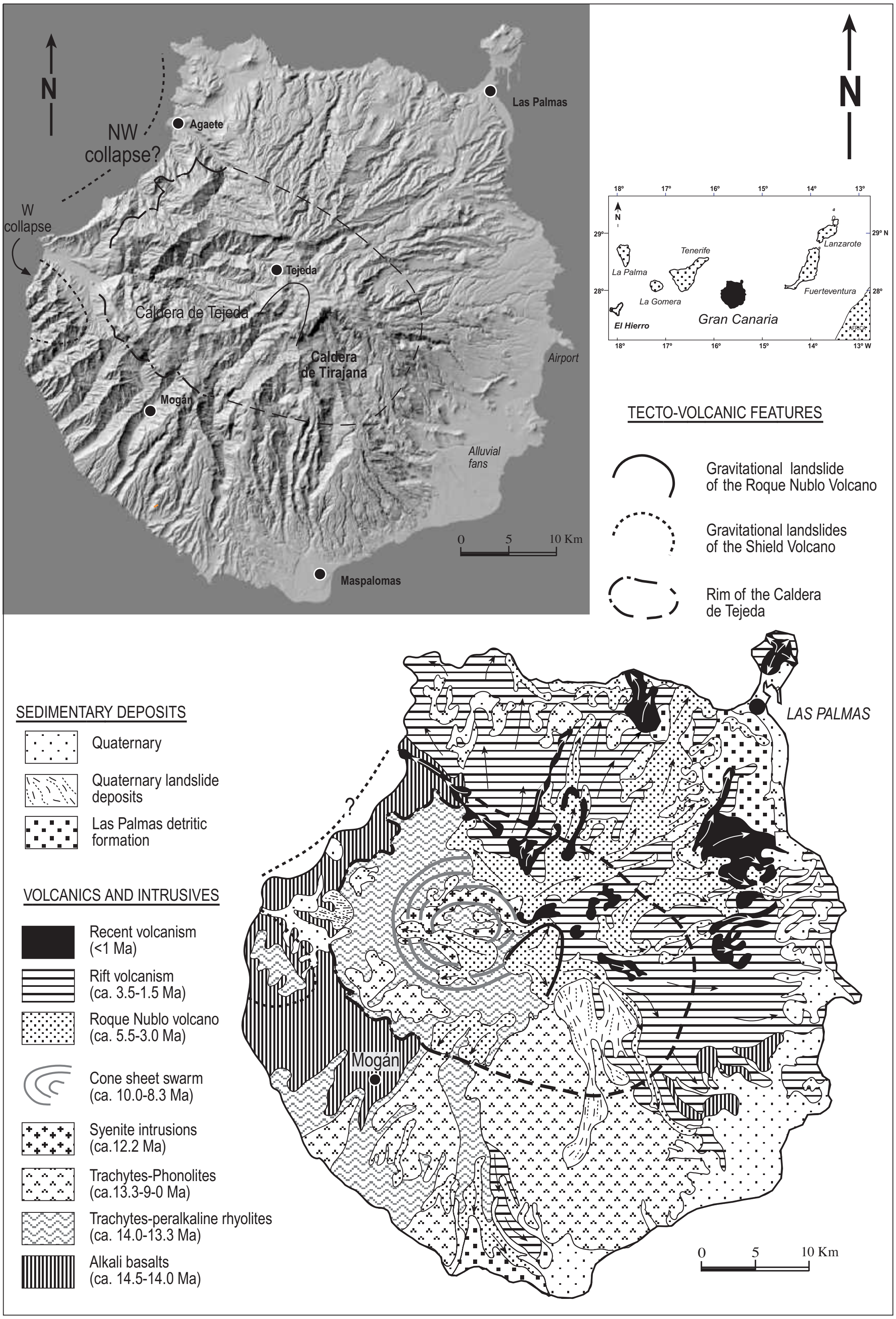

Fig. 18.5. Simplified geological map of Gran Canaria (modified from ITGE 1992). Inset: shaded relief image of Gran Canaria indicating the main geomorphological and tectonic features (image GRAFCAN).

as well as several kilometres offshore to the west (Schmincke 1976, 1993; ITGE 1990, 1992). At present, shield-stage rocks are exposed mainly in the western and southwestern coastal cliffs, where they reach thicknesses of up to 1000 m (Fig. 18.5). Volcanic activity was characterized by Hawaiian-type eruptive styles, involving sustained eruption of effusive lavas (alkali-basalts to mugearites) with minor interlayered pyroclastics and a swarm of feeder dykes (Fúster *et al.* 1968*d*; Schmincke 1976, 1993). The growth period for this stage is very short, constrained to between 14.5 to 14 Ma BP according to K-Ar and $^{39}Ar/^{40}Ar$ ages from Bogaard *et al.* (1988) and Bogaard & Schmincke (1998). This short time of growth and the sparse occurrence of pyroclastic rocks strongly contrasts with the shield stages of the other Canaries, where shield stages exceeded at least 1 Ma and produced predominantly pyroclastic facies from more explosive eruptions.

A clear intraformational unconformity at the SW of the island suggests the presence of a gravitational collapse embayment, with thick, interbedded debris-avalanche deposits and infilling lavas of slightly younger ages mass wasting the southwestern flank of the shield (Schmincke 1976, 1993). The arcuate outline of the NW coast (San Nicolás to Agaete) has also been attributed to a gravitational collapse. The original shape and extent of the volcanoes forming the shield stage of Gran Canaria are unknown, although Schmincke (1976, 1993) defined three volcanoes: the northwest volcano, centred close to Agaete; the north volcano, near the Aldea de San Nicolás, and the southeast volcano, with a centre close to Agüimes. Schmincke & Sumita (1998) postulated a fourth volcano to the north of Arucas, and suggested an east to west migration of volcanic activity in the shield stage of Gran Canaria, with the southeast volcano being the youngest. Analysis of dyke distribution and orientation carried out during the geological mapping of the island (ITGE 1990, 1992) suggests the presence of a main shield volcano centred in the Mesa del Junquillo (at the northern edge of the cone sheet west of Tejeda). The volcanoes defined by Schmincke (1976, 1993) and Schmincke & Sumita (1998) may have functioned as rift zones associated with this main shield volcano, which subsequently collapsed to form the Caldera de Tejeda.

Caldera stage volcanism

A shallow (*c.* 4–5 km) rhyolitic magmatic chamber was emplaced towards the final stages of shield development, periodically fed from a deeper (sublithospheric *c.* 14 km) basaltic source (Freundt & Schmincke 1992). These rhyolitic magmas produced the first highly explosive eruptions and, subsequently, highly welded ignimbrites. Ignimbritic eruptions and high eruptive discharges characteristic of the shield stage may have caused the abrupt emptying of the magmatic chamber, finally collapsing the summit of the volcano and creating the Caldera de Tejeda (Schmincke 1967; Hernán 1976). This elliptical, NW–SE trending, 20 × 17 km, 1000 m deep caldera is the most prominent geomorphological feature of Gran Canaria (Fig. 18.5). The caldera rim, with contacts usually dipping 45° towards the centre of the island, is observable at present along its western half, and is marked by bright green-blue hydrothermal alteration layers (locally known as ‘azulejos’). Simultaneously with the formation of the caldera, the rhyolitic chamber was replenished with basaltic magmas. The weight of the cauldron block forced the violent emission of *c.* 80 km^3 of magma through the caldera rim fissures (Freundt & Schmincke 1992), producing a single, 30 m thick ignimbrite named P1 by Schmincke (1976, 1993) and dated at 14 Ma BP (Bogaard *et al.* 1988; Bogaard & Schmincke 1998). The P1 ignimbrite covers >400 km^2 of the shield-stage basalts around the caldera and constitutes a prominent marker bed recording the creation of the Caldera de Tejeda.

Post-caldera stage volcanism

This stage is characterized by the eruption of large volumes (*c.* 1000 km^3) of silicic lavas and ignimbrites from the caldera rim (ring fractures). The presence of ‘fiamme’ and rheomorphic folds suggests boiling-over-type eruptive mechanisms during expulsion of high temperature ignimbrites. Two magmatic phases, comprising extra- and intra-caldera domains, can be defined: a predominantly peralkaline rhyolite–trachyte initial phase (Mogán Group (Gp) in the terminology of Schmincke (1976, 1993)) and a later trachyte–phonolite phase (Fataga Gp). Both phases were fed from a shallow magma reservoir, periodically replenished with deeper basaltic magmas, later differentiated by crystal fractionation (Schmincke 1976, 1993). A peralkaline rhyolite–trachyte initial phase (*c.* 14–13.3 Ma BP; volume estimates of about 300–500 km^3) was heralded by the eruption of the P1 ignimbrite. The extra-caldera deposits, composed of 15 cooling units up to 300 m thick (Schmincke 1976, 1993), mantle most of the shield volcano. Flow direction and periclinal slopes dipping at 7–9° relate these eruptions to the caldera rim (Schmincke & Swanson 1967). The lack of interbedded epiclastic deposits suggests high eruptive rates and short periods of emission. The intra-caldera volcanic deposits filled most of the caldera, but later intrusions have hindered stratigraphic correlation with the extra-caldera volcanics. However, it seems evident that both were erupted from the same vents located at the edge of the caldera.

During the trachyte–phonolite second eruptive phase (*c.* 13.3–8.3 Ma BP; volume estimates >500 km^3), extra-caldera deposits formed sequences up to 1000 m thick of successive ignimbrites with interlayered lava flows that become more abundant towards the top. Interbedded epiclastic deposits indicate frequent interruptions of eruptive activity and episodic mass wasting processes, determined by Bogaard & Schmincke (1998) to have occurred between 12.33–12.07, 11.36–10.97 and 9.85–8.84 Ma BP. Eruptions continued from vents along the caldera rim and, probably, from a strato-volcano located near the Cruz Grande, in the SE of the island (Schmincke 1976, 1993).

Intra-caldera activity in this second phase is mainly intrusive, consisting of three main events (Schmincke 1967, 1976, 1993; Hernán 1976): alkali syenites (*c.* 12.2 Ma BP), trachytic-phonolitic cone sheets (*c.* 10–8.3 Ma BP), and phonolitic-nephelinitic plugs (*c.* 8.5 Ma BP). The syenites form small stocks located at altitudes up to 1200 m in the centre of the caldera, whereas the phonolitic-nephelinitic plugs define a circular trend at the edge of the 12 km diameter cone sheet outcrop. Trachytic and phonolitic intrusions cut the complete sequence, producing a 3 km doming of the region (Hernán & Vélez 1980). Dyke density increases sharply northwestards to form >90% of the outcrop, and dyke dips increase from about 30° on the periphery to 50° in the centre. In cross-section, the cone-sheet distribution points to a common focus located at a depth of about 2 km (Hernán & Vélez 1980). The similarity in age and composition of all these intrusives with the extra-caldera trachyte-phonolite ignimbrites and lavas suggests that they represent a subvolcanic facies.

Erosional stage

Following the shield and caldera stages the island entered a long period (*c.* 3 Ma) of erosion, with eruptions limited to minor phonolitic events on the northern slopes (Pérez Torrado *et al.* 2000). A radial pattern of canyons cut deep into the island, reaching the basaltic shield volcano. This palaeotopography controlled the distribution of the subsequent post-erosional volcanism (Schmincke 1976, 1993).

Sediments originated in this erosional period were deposited in alluvial fans predominantly on the northeastern, eastern, and southern coastal platforms (Fig. 18.5), forming a Lower Member (Mb) of the Las Palmas Detrital Fm (LPDF; ITGE 1990, 1992). The onset of renewed volcanism (see below) coincides in Gran Canaria with an important transgression (Lietz & Schmincke 1975), during which a Middle Mb of the LPDF (ITGE 1990, 1992) was formed. These marine sediments, with abundant Pliocene fauna, now crop out in the NNE at heights of 50–110 m (ITGE 1990).

Post-erosional magmatism

Renewed volcanism (5.5 Ma BP to present), has shown three main phases: Roque Nublo, post-Roque Nublo and recent volcanism (Pérez Torrado 2000). Roque Nublo volcanism involved strombolian eruptions of basanitic and nephelinitic magmas localized in the central and southern slopes of the island, initially forming a NW–SE aligned series of vents. Later eruptions (4.6 Ma BP) became focused in the centre of the island, building the large and complex Roque Nublo stratovolcano over a period of >1.5 Ma (Pérez Torrado *et al.* 1995). Volcanic activity began with the emission of large amounts of lavas (basanites–alkali basalts to trachytes–phonolites) that were immediately channelled into the network of canyons excavated during the previous erosional stage. Some of the first mafic lavas flowed up to 20 km towards the NNE coastal areas, where they formed a thick (>20 m) sequence of pillow lava flows, with minor pillow breccia and hyaloclastites, above the marine deposits of the Middle Mb of the LPDF. When the magma changed to trachytic-phonolitic compositions (at about 3.9 Ma BP), explosive volcanic activity occurred in the summit area of the stratovolcano, producing breccia-type ignimbritic deposits (Pérez Torrado *et al.* 1997). Eruptive activity of the volcano ended by around 3 Ma BP with the intrusion of phonolitic plugs (see Fig. 18.18B).

The distribution and geometry of the Roque Nublo volcanics (*c.* 200 km^3) suggest that this volcano may have exceeded 2500 m in height, with asymmetric flanks defined by extended and gentle slopes in the north and short and steep sides in the south (Pérez Torrado *et al.* 1995). Several gravitational collapses mass wasted the Roque Nublo stratovolcano, generating 25 km long debris avalanche deposits (García Cacho *et al.* 1994; Mehl & Schmincke 1999). These deposits have been identified in the submarine ODP Leg 157 boreholes (Schmincke & Sumita 1998).

The second main magmatic phase (Post-Roque Nublo) has been thought by some authors to have started after a >0.5 Ma gap in volcanic activity (McDougall & Schmincke 1976; Schmincke 1976, 1993; ITGE 1990, 1992). However, the Roque Nublo terminal phonolitic intrusions seem coeval with the initial basanitic eruptions of the post-Roque Nublo phase (Pérez Torrado *et al.* 1995) and the epiclastic deposits considered representative of this erosive gap have been subsequently reassigned to different stratigraphic units of the island. The volcanic activity of this stage is characterized by strombolian vents along a NW–SE rift, with basanite–nephelinite and trachybasaltic lavas forming a 500 m thick sequence (about 10 km^3) that covers large areas on the northern slopes of the island. The ages of these lavas suggest the main eruptive period was from 3 to 1.7 Ma BP (McDougall & Schmincke 1976; ITGE 1990, 1992).

Finally, the most recent volcanism has involved rare, dispersed minor eruptions of highly alkalic magmas (basanites, nephelinites). These eruptions produced phreatomagmatic tuff cones and calderas such as the Caldera de Bandama, and strombolian vents such as the Montañón Negro volcano, dated at 3.5 ka BP (Nogales & Schmincke 1969), one of the youngest eruptions known on Gran Canaria.

La Gomera (FH, CRC)

La Gomera is intermediate in size (*c.* 380 km^2) between the two other western islands, La Palma and El Hierro (Fig. 18.1). The island reaches a maximum height of 1487 m in the centre (Alto de Garajonay) above a small (40 km^2) plateau ('meseta') from which several radial, deep, and narrow ravines run down to the coast (Fig. 18. 6 inset). These ravines interrupt steep coastal cliffs, 50 to 850 m high, and those in the south are separated by gently sloping flat-topped ridges, which form a distinct feature of the island. Other noteworthy features of La Gomera landscape include the common presence of the so-called 'roques' or 'fortalezas', remains of eroded felsic lava plugs and domes, and wall-like subvertical dykes which stand out from the host rock because of preferential erosion. By contrast, due to the lack of Quaternary volcanic activity, recent cinder cones and lava fields, very common on all the other islands, are absent on La Gomera. The only well-preserved volcanic cone is La Caldera, near the southern coast (Fig. 18.6).

The geology of this island can be subdivided into an old basement of submarine and intrusive origin, an overlying trachytic-phonolitic complex, older and younger basalts, and a series of late felsic domes (Fig. 18.6). Previous studies include those of Gagel (1925), Blumenthal (1961), Bravo (1964) and Hausen (1971), with more specific publications dealing with the old basement (Cendrero 1971), geochronology (Abdel-Monem *et al.* 1971, 1972; Féraud 1981; Cantagrel *et al.* 1984), felsic domes (Cubas 1978), and trachytic-phonolitic rocks (Rodríguez 1988).

The oldest rocks exposed on La Gomera are known as the basal complex and represent the basement developed during the pre-shield stage of the island (Cendrero 1971). As in Fuerteventura and La Palma, submarine (basaltic and trachytic) pillow lavas and tuffs and very thinly bedded (pelitic, silicic, and carbonate) sediments crop out on La Gomera, but they are restricted to a few small exposures and some screens in the north, and so there is a lack of palaeontological age data. The main rocks forming this La Gomera basal complex are mafic (gabbros and olivine gabbros) and ultramafic (wehrlites and pyroxenites) in composition, and as extremely dense dyke swarm which represents up to 90% of the unit. Radiometric (K-Ar) ages for intrusive lithologies range between 20 and 14 Ma BP, and probably record late stages of cooling and

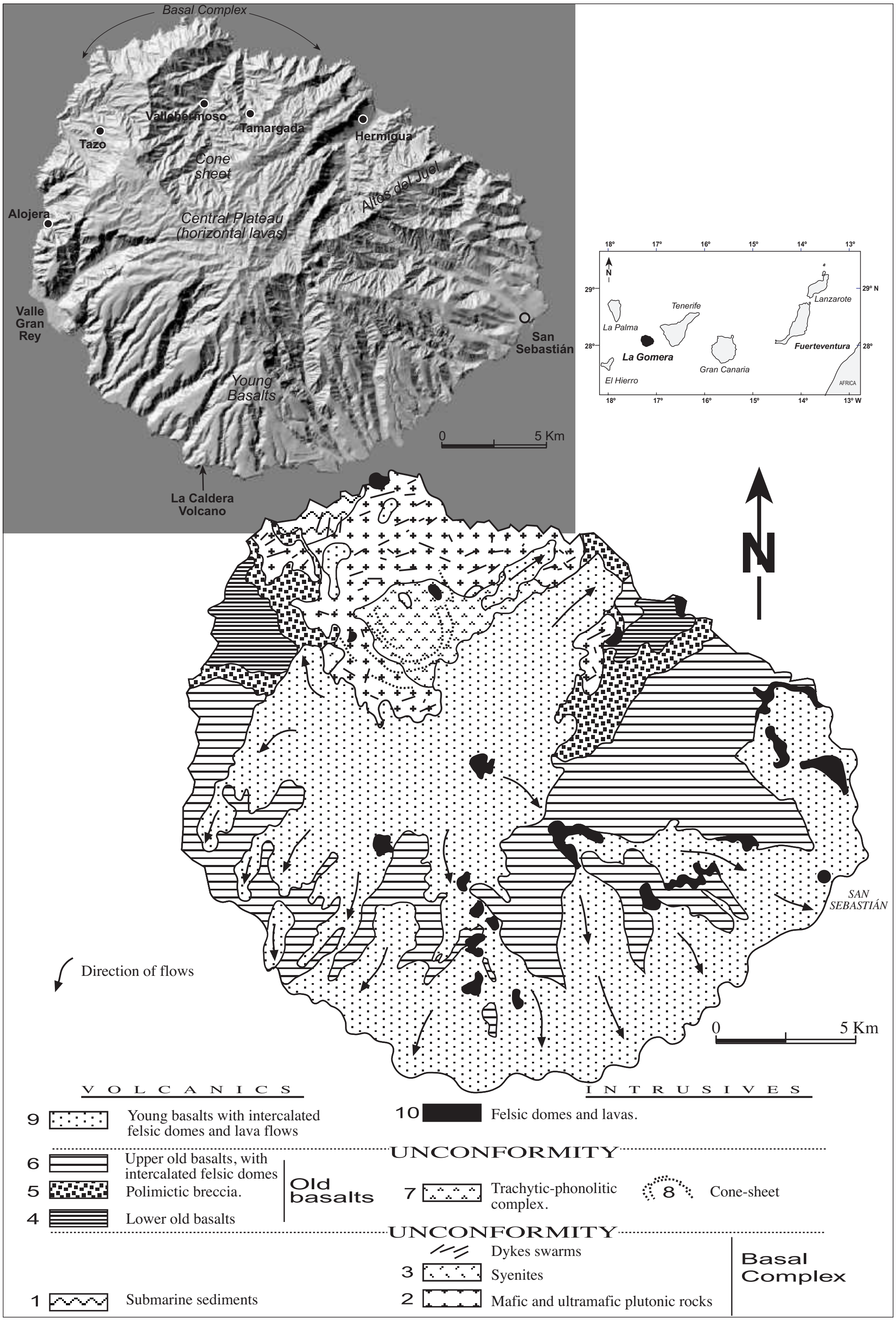

Fig. 18.6. Simplified geological map of La Gomera. Inset: shaded relief image of La Gomera with indication of the main geomorphological and tectonic features (image GRAFCAN).

emplacement. The only outcrop of syenite in the island (usually considered to be part of the basal complex) is to be found in Tamargada. These fine-grained rocks are clearly the youngest among the plutonics and therefore yielded a much younger K-Ar age of *c.* 9 Ma (Cantagrel *et al.* 1984). Most of the dykes are basaltic in composition, but show a great variety of aphanitic and porphyritic types that give rise to a complicated, irregular network. Several felsic (trachytic to phonolitic) dykes are also present, but these are related to the overlying trachytic-phonolitic series.

The trachyte–phonolite complex, recording the next phase in the magmatic evolution of the island, crops out in a deeply eroded, well drained area between Las Rosas and Vallehermoso. It comprises fractured and highly weathered trachytic to phonolitic massive lavas, breccias, domes and dykes that seem to lie unconformably on the basement, although the contact relationship is rarely exposed. The dykes, which are abundant but not so numerous as in the basal complex, have been grouped into two radial swarms (Huertas *et al.* 2000) and one cone sheet complex (Rodríguez 1987; Hernán *et al.* 2000). The earlier radial swarm, the axis of which is centred at Tamargada, is most likely related to the syenites exposed in the same area, whilst the later one, centred south of Vallehermoso, seems to be associated with the cone sheets which were fed from a dome-shaped magma body some 1350 m below sea level (Hernán *et al.* 2000).

A group of basaltic rocks known as the 'lower old basalts' lies unconformably on the basal complex, but, unfortunately, a lack of critical exposures where these basalts are observed in direct contact with the trachytic-phonolitic rocks has led to questions about their relative stratigraphic position. Some authors have considered the trachytic-phonolitic complex to be the oldest subaerial unit, whereas others contend that the lower old basalts represent the first subaerial activity. A few felsic dykes, probably belonging to the trachyte–phonolite complex, cut through these lower old basalts. If the latter interpretation is accepted, then these old basalts mark the onset of the island shield stage. The unit consists mainly of thin pahoehoe lava flows, and subordinate interlayered pyroclastic horizons, cut by a dense dyke swarm, with abundant sills. Most lava flows are porphyritic ankaramites and plagioclase basalts that reach a maximum thickness of 250 m in the single large area where they are exposed, between Tazo and Alojera in the NW. Smaller outcrops include one near the town of Hermigua, where Cubas *et al.* (1994) identified submarine levels at the base of the sequence, and one or two more restricted spots in the south where the barrancos have excavated deeper. These outcrops represent remnants of a larger edifice that once occupied the northern onland and offshore areas of the present island. The initial subaerial basaltic shield may have been active between at least 11 and 9 Ma BP, as suggested by the K-Ar age from a dyke (Cantagrel *et al.* 1984) and two ages from lava flows near Hermigua (Cubas *et al.* 1994).

A polymictic breccia up to 300 m thick lies unconformably over the lower old basalts. This breccia includes heterogeneous, subrounded and angular clasts up to 50 cm in diameter and surrounded by a cemented fine-grained groundmass. Its origin has been interpreted as due to either highly explosive or laharic activity. Occasionally, where the lower old basalts are lacking, the breccia rests directly over the basal complex rocks. The breccia is presumably a remnant of an almost entirely removed initial shield volcano. Near its top, the breccia is conformably interlayered with new basaltic flows known as the 'upper old basalts'. These new basalts are spread over the entire island and are better preserved than the lower old basalts. They were erupted effusively from fissure vents and built a large new shield, which has since been capped by younger rocks. The remains of this shield are at present exposed by erosion at the bottom and in the walls of the main barrancos, with the maximum visible thickness (over 500 m) being preserved in Altos de Juel. The pile consists mainly of thin lava flows (ankaramites, plagioclase basalts, and aphanitic basalts), some pyroclastic layers and a few buried pyroclastic cones, with thick trachybasalt flows near the top of the sequence. According to Cubas (1978), a few of the trachytic domes cutting only the upper old basalts but not the young basalts (see below) could also be a part of this unit. However, other authors (Bravo 1964; Cendrero 1971; Rodríguez 1988) distinguished a single, more recent episode of felsic domes. The numerous dykes intruding the old basalts, more abundant in the lower than in the upper ones, acted as feeders for the different subaerial units. The most consistent radiometric (K-Ar and $^{39}Ar/^{40}Ar$) ages obtained by several authors (Abdel-Monem *et al.* 1971; Féraud 1981; Cantagrel *et al.* 1984) place the upper old basalts activity between about 9 and 6.5 Ma BP. The available age data suggest that either an interval of quiescence accompanied by gradual denudation or a violent rapid destruction of the lower basaltic shield took place between approximately 10 and 9 Ma BP.

The most recent basaltic unit is separated from the upper older basalts by a marked unconformity representing *c.* 1.5 Ma and clearly visible only in the western area. The unit comprises a thick (500 to 1000 m) pile of mostly basaltic lava flows in which olivine and augite-bearing porphyritic basalts, aphanitic basalts and trachybasalts are the most representative types. Plagioclase basalts, common in earlier episodes, were only rarely erupted during this younger activity. The individual flows are very thick (5–10 m), only rarely cut by dykes, and generally well preserved, exhibiting frequent columnar, and sometimes spherical jointing. Whilst in the central sector the flows are characteristically flat, they dip gently seawards in the southern peripheral sector. Basaltic pyroclasts are fairly well represented in the form of interlayered sheets and buried cones, and felsic pyroclastic layers, lava flows and, most of all, phonolitic domes also occur. The previously mentioned cone, La Caldera, belongs to this phase of activity, as does another, less well-preserved one almost completely covered by vegetation in the central sector. K-Ar and $^{39}Ar/^{40}Ar$ ages determined from basaltic flows and felsic domes indicate concentrated activity approximately between 5 and 4 Ma BP, although a single datum from a lava at the top of the section in Arure would extend the activity up to 2.8 Ma BP. The emission of the latest, more heterogeneous materials must have accentuated the island relief and enlarged its surface considerably. How large and high the island became after the construction of this third and most complex shield is not known. High coastal cliffs around the island, and especially along the north and west coasts, suggest that the present island is only a small fraction of the size it once was.

Tenerife (EA)

Tenerife is the largest (2058 km^2) and highest (3718 m) island in the Canaries and has a complex volcanic history. The oldest visible materials (the Old Basaltic Series; Fúster *et al.* 1968*a*) are preserved in three isolated and deeply eroded massifs (Fig. 18.7): Anaga (to the NE), Teno (to the NW) and Roque del Conde (to the south). Each has been formed in late Miocene and early Pliocene times as a result of several independent volcanic cycles interspersed with long pauses in activity (Ancochea *et al.* 1990).

After the Old Basaltic Series, the volcanic activity became concentrated within two large edifices: the central composite volcano of Las Cañadas and the 'cordillera dorsal', a SW–NE volcanic ridge linking Las Cañadas and the Anaga massif (Fig. 18.7). On each side of the cordillera dorsal, two large trapezoidal depressions have been formed, the so-called 'valleys' of Güímar and La Orotava.

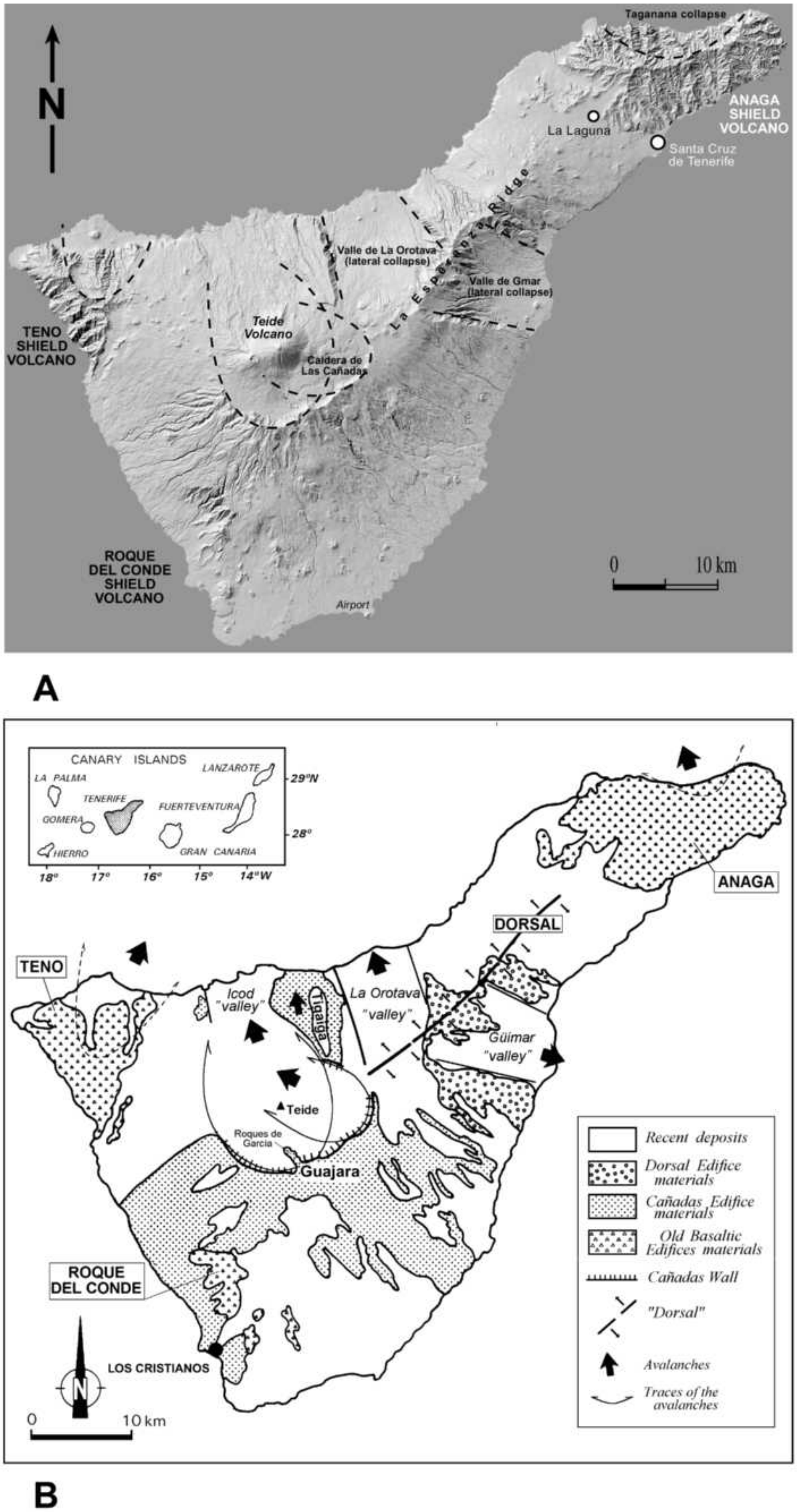

Fig. 18.7. **(A)** Shaded relief image of Tenerife with indication of the main geomorphological and tectonic features (image GRAFCAN). **(B)** General sketch map of Tenerife (modified from Cantagrel *et al.* 1999 and Ancochea *et al.* 2000).

On the upper part of the Las Cañadas volcanic edifice, there is a semi-elliptic depression with a NE trending axis 16 km long, known as the 'Caldera de Las Cañadas'. This depression is partially surrounded by a southeastern cliff, Las Cañadas Wall, over 25 km long and reaching a maximum altitude (2712 m) and height (500 m) at Guajara (Fig. 18.7). The most recent activity is represented by many monogenetic basaltic centres scattered across the island and by the Teide–Pico Viejo edifice, volcanic products from which partially fill the Las Cañadas caldera. Some basaltic eruptions have taken place in historical times (last 500 years). Further details and additional references on the geology of the island can be found in Fúster *et al.* (1968*a*), Carracedo (1979) and Ancochea *et al.* (1990, 1999).

With regard to the oldest massifs, that of Anaga comprises a complex sequence of basaltic lava flows and volcaniclastic levels intruded by basaltic and phonolitic dykes and domes (Fig. 18.8a). Three units have been identified (Fig. 18.9), the oldest of which (Lower Anaga) appears in the north in a large arcuate landform called Arco de Taganana and consists of volcaniclastic rocks and debris flow deposits cut by abundant dykes. The age of this unit is not well defined, due to intense alteration and the fact that a K-Ar age of 16.1 Ma from an ankaramite of this unit (Abdel-Monem *et al.* 1972) is much older than any other for Tenerife and so must be viewed with caution (Ancochea *et al.* 1990). The overlying Middle Anaga unit has a total thickness of about 1000 m, is separated from the former by an unconformity, and dips towards the south. It comprises basaltic pyroclastic and lava flows, with subordinate felsic plugs and lava flows. The ages obtained from this unit vary between 6.5 Ma and 4.5 Ma (Carracedo 1975; Féraud 1981; Féraud *et al.* 1985; Ancochea *et al.* 1990). The youngest unit (Upper Anaga: 3.3–3.7 Ma BP) is basaltic to phonolitic in composition, appears in the western part of the massif, and lies above a clear unconformity.

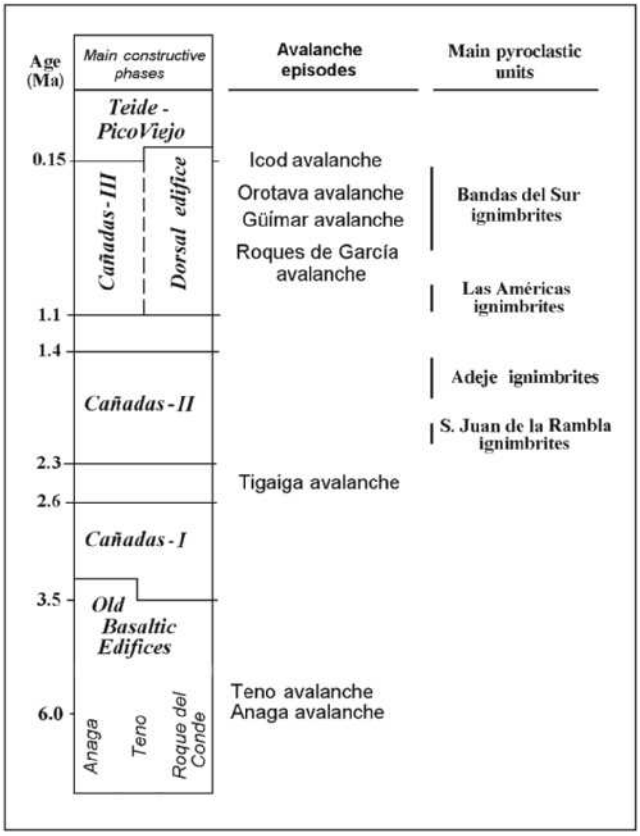

Fig. 18.9. Simplified volcanic stratigraphy, avalanche episodes and main pyroclastic units of Tenerife (modifies from Cantagrel *et al.* 1999).

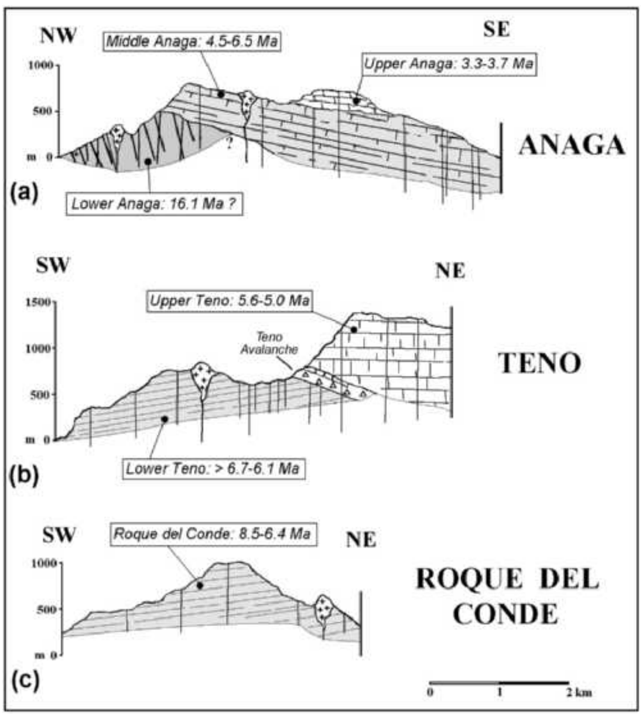

Fig. 18.8. Schematic cross-sections of old basaltic edifices in Tenerife.

The Teno edifice is made up of only two units: a lower sequence (Lower Teno) of basaltic pyroclastics and lava flows, dipping seawards, covered unconformably by an upper sequence (Upper Teno) of subhorizontal (700 m) basaltic lava flows, with some trachytes, and cut by abundant dykes (Fig. 18.8b). The upper part of the Lower Teno unit has been dated by Ancochea *et al.* (1990) at 6.7–6.1 Ma BP, the Upper Teno unit at 5.6–5.0 Ma BP, and the youngest age yet obtained is from a 4.5 Ma phonolitic plug. In the lower part of the Upper Teno unit a debris flow deposit (*c.* 6 Ma BP; Cantagrel *et al.* 1999) represents the subaerial remains of a major collapse (Teno avalanche; Fig. 18.8b). Finally, the Roque del Conde edifice comprises 1000 m of basaltic lava flows (Fig. 18.8c), with ages mainly between 8.5 and 6.4 Ma (Ancochea *et al.* 1990).

Over the last 3.5 Ma the central part of Tenerife has been occupied by shield or central composite volcanoes that at times have reached more than 3000 m in height (Ancochea *et al.* 1999, 2000). Four main phases have been recognized: Cañadas I, II, III, and the presently active Teide–Pico Viejo complex ('Cañadas IV'; Fig. 18.9). Initial volcanic activity (Cañadas I volcano) took place in the western part of the island, between *c.* 3.5 Ma and 2.6 Ma BP, and produced mainly basalts, trachybasalts and trachytes. The remains of this phase crop out in the Cañadas wall and at the bottom of several radial ravines. Its main emission centre (3000 m high) was located in the central part of what is now the Cañadas 'caldera' (Ancochea *et al.* 1999). This edifice underwent partial destruction by failure and flank collapse to the north (Ibarrola *et al.* 1993; Cantagrel *et al.* 1999) forming a debris avalanche deposit (Tigaiga breccia) on which a second volcano was built (Cañadas II volcano) between 2.4 and 1.4 Ma BP. Towards the end of this second period of activity, major explosive eruptions took place forming ignimbrites, pyroclastic flows, and ash fall deposits of trachytic composition, now exposed in the SW lower slopes.

The next constructional phase, from 1.1 to 0.15 Ma, built a new volcano (Cañadas III volcano) which produced trachybasaltic lava flows and abundant phonolitic products. The later stages (0.7–0.15 Ma BP) are again strongly marked by major explosive eruptions, producing the Bandas del Sur ignimbrites from an eastern eruptive centre, whose pyroclastic products are widely exposed across the SE lower slopes of Tenerife. Several of these pyroclastic eruptions had sufficient volume to trigger the formation of a complex vertical collapse caldera(s) in the summit region (Martí *et al.* 1994, 1997; Bryan *et al.* 1998). However, their geometry and location within the Cañadas II and III edifices is poorly defined because several debris avalanche events later broke through these earlier structures.

The formation mechanisms of the present Las Cañadas 'caldera' have long been discussed. Models for erosion, collapse, explosion and avalanche have been proposed (e.g. Hausen 1949, 1956, 1961; Bravo 1962; McFarlane & Ridley 1969; Araña 1971; Coello 1973; Booth 1973; Ancochea *et al.* 1990, 1998; Carracedo 1994; Martí *et al.* 1994; Watts & Masson 1995). In recent years, the discussion has focused on the alternatives of vertical collapse versus lateral collapse, or a combination of both. Martí *et al.* (1997) suggested that Las Cañadas is the result of a complex sequence of vertical collapse events. These vertical collapses may have played a major role in triggering lateral collapses. For Cantagrel *et al.* (1999) the present shape of the wall at Las Cañadas is not related to vertical collapses, but is instead a product of repeated flank failures.

Numerous pyroclastic trachytic and phonolitic eruptions occurred during the construction of Las Cañadas, the early studies of which provided a landmark in the international development of tephrochronology (Walker 1973; Booth 1973). This pyroclastic activity showed an eastward migration through time (Martí *et al.* 1994; Ancochea *et al.* 1999) and became increasingly common, so that pyroclastic deposits of all types represent a large proportion of the volcanic products that make up the Cañadas volcanic edifice (Ancochea *et al.* 1990, 1999; Martí *et al.* 1994). Pyroclastic rocks are especially well exposed on the SE lower slopes (Bandas del Sur) of the island, where they have been extensively studied (Alonso 1989; Bryan *et al.* 1998), but are also present in various stratigraphic positions in all the other sectors of the Cañadas edifice. According to their geographic position and isotopic ages, four main pyroclastic phases may be identified in the Cañadas edifice over the last 2 Ma (Figs 18.9 and 18.10). The first 'San Juan de la Rambla' phase (*c.* 2 Ma BP), whose outcrops are presently restricted to the north of Tenerife, occurred during the first period of construction of the Cañadas II edifice. The second 'Adeje' phase (1.8–1.5 Ma BP) is much more important in volume and erupted in successive distinct pyroclastic events during the second part of construction of the Cañadas II edifice. Widespread in the SW, the pyroclastic

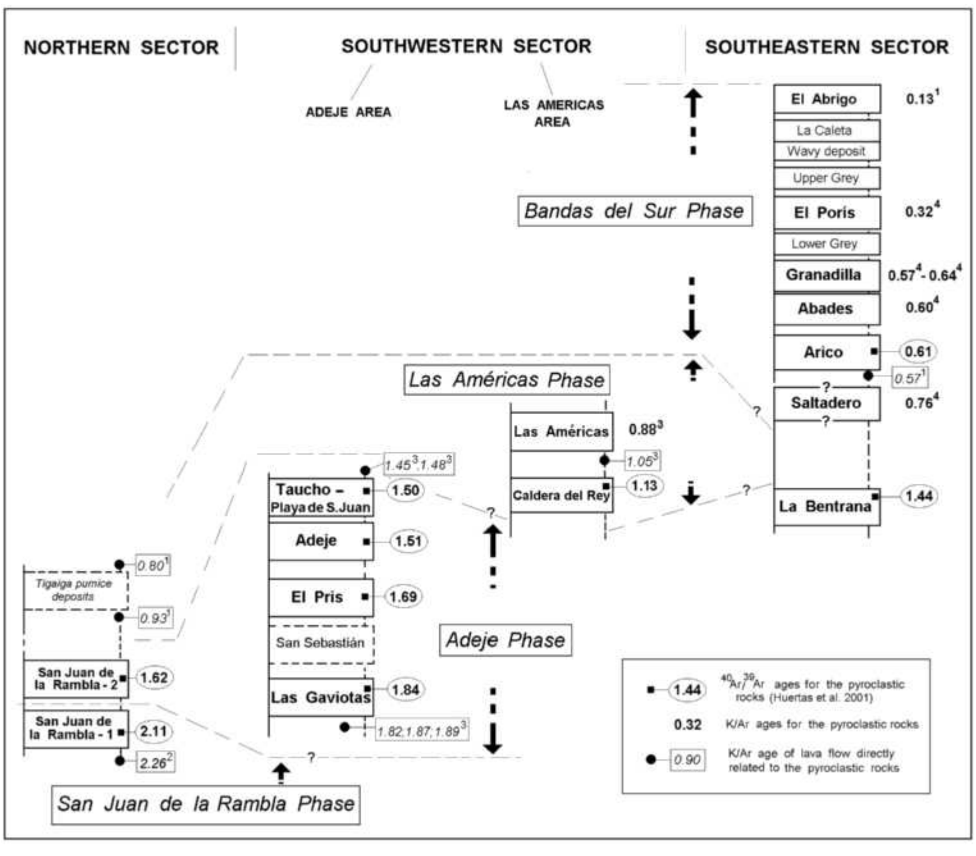

Fig. 18.10. Schematic stratigraphic colums of the main pyroclastic deposits showing possible correlations. K-Ar ages (in Ma) from the following references: [1]Ancochea *et al.* (1990); [2]Ibarrola *et al.* (1993); [3]Fúser *et al.* (1994); [4]Bryan *et al.* (1998).

deposits of this phase have also been recognized in the Tigaiga massif in the north and in the deepest levels of the Barranco de la Bentrana in the SE (Bandas del Sur). The third 'Las Américas' phase (1.1–0.9 Ma BP) is presently recognized around the town of Los Cristianos, in the south of the island, and in the Tigaiga massif in the north, although some pyroclastic units stratigraphically located between the La Bentrana and Arico ignimbrites could also belong to this phase (Fig. 18.10). The fourth 'Bandas del Sur' phase (Wolff 1985; Alonso 1989; Bryan *et al.* 1998; 0.7–0.15 Ma BP) is by far the best exposed and also appears to be the most important in volume. Widespread across the SE slopes of Tenerife, it is known in the north at the top of the Tigaiga massif but has not yet been described in the SW. Several tens of pyroclastic units have been identified but their lateral extent is commonly limited so that no unique stratigraphic column can be drawn, although a simplified representation is presented in Fig. 18.10 (see Bryan 1995 and Bryan *et al.* 1998 for details). The coeval eruption of the dorsal edifice (Ancochea *et al.* 1999) produced a basaltic NE-trending volcanic ridge, the westernmost flows of which are interbedded with felsic rocks from the Cañadas III central edifice. The so-called 'valleys' of La Orotava and Güímar, transversal to the ridge axis (Fig. 18.7) also formed during this period. Finally, the most recent volcanic activity in Tenerife is represented by the Teide–Pico Viejo complex, formed by basalts, trachytes and phonolites, which partly fill Las Cañadas 'caldera' (e.g. Ablay *et al.* 1998), and by many small scattered basaltic monogenetic volcanoes, mainly on the dorsal ridge. Six such basaltic volcanoes have erupted in historical times (see Fig. 18.21) and the Teide–Pico Viejo complex has also shown historical activity.

Much attention has recently focused on very large debris avalanche deposits found off the north coast of Tenerife (Ancochea *et al.* 1990, 1998; Carracedo 1994; Watts & Masson 1995; Teide Group 1997). On the SE flank of the dorsal ridge, a south-directed, large landslide formed the Güímar valley later than 0.80 Ma BP. Another six, north-directed debris avalanche events have been identified in Tenerife (Cantagrel *et al.* 1999): the Anaga and Teno (*c.* 6 Ma BP) in the Old Basaltic edifices, followed by the Tigaiga event in Cañadas I volcano (>2.3 Ma BP), Los Roques de García (<1.4 Ma BP); La Orotava avalanche (*c.* 0.6 Ma BP) that formed the Orotava valley; and the <0.15 Ma BP Icod avalanche (Fig. 18.10).

El Hierro (JCC, FJPT, ERB)

The island of El Hierro, SW of La Palma (Fig. 18.1), is the smallest (287 km^2) and least populated of the archipelago. It is also one of the steepest oceanic shield volcanoes in the world, with slopes that frequently exceed 30°, with the emergent summit, Malpaso, rising to 1501 m from a depth below sea level of 3700–4000 m. The characteristic trilobate morphology of the island is the result of the development of a regular (120°), three-branched rift zone (Carracedo 1994, 1996*a*) that forms the ridges of the volcano (Fig. 18.11). The sectors between the rifts form arcuate embayments, originating by catastrophic mass wasting during successive giant landslides. El Hierro and La Palma form a dual line of island volcanoes, with volcanism apparently alternating between the two islands (Carracedo *et al.* 1999*a*), and only La Palma showing important eruptive activity during Holocene times.

Published reports on the geology of El Hierro are relatively scarce prior to 1996 and focused on the general geology and petrology of the island (Hausen 1964), and on the geochronology (Abdel-Monem *et al.* 1972; Fúster *et al.* 1993). However, after 1996, much work has been published on the geochronology (Guillou *et al.* 1996), geology and tectonics (Day *et al.* 1997; Carracedo *et al.* 1999b) and submarine geology (Masson 1996; Urgelés *et al.* 1997, 1998). Recent mapping with K-Ar dated geomagnetic reversals has made possible stratigraphic correlation between selected sections, permitting the reconstruction of the geological evolution of the island and the definition of three main volcanic edifices (Fig. 18.11): the Tiñor and El Golfo volcanoes, and the subsequent development of a triple rift system.

Tiñor volcano

This constitutes the first stage of subaerial growth of El Hierro and its present outcrop is confined to the NE flank of the island and the interior of the Las Playas embayment (Fig. 18.11). Tiñor volcano lavas are basaltic and characterized geochemically by relatively primitive picrobasaltic to hawaiitic-tephritic compositions. According to the K-Ar ages and magnetostratigraphy, Tiñor volcano developed very rapidly and continuously between *c.* 1.2–0.8 Ma BP. There is no consistent compositional variation with time that can be mapped in the field, but it is possible to recognize early, intermediate and late stage magmatic phases. The early stage has produced a basal unit of thin flows on steep slopes, with more pyroclastic rocks and dykes than in younger basalts. The intermediate stage shows thicker lavas with dips that become subhorizontal in the centre of the edifice, probably reflecting the mature stage of growth of the volcano. Late stage activity produced more explosive, wide-cratered vents (the Ventejís volcano group) and associated xenolith-rich lavas. This explosive stage seems to have immediately preceded the collapse of the NW flank of the Tiñor volcano at *c.* 0.88 Ma BP (Figs 18.11 and 18.12).

El Golfo

This represents a new volcanic edifice developed over the remnants of the Tiñor volcano, filling its collapse embayment and totally burying it. Continued volcanism from *c.* 0.55 to 0.17 Ma BP built a 2000 m high shield volcano, 20 km in diameter, with a basal, mostly pyroclastic unit overlain by an upper unit composed mostly of lava flows. The basal unit was produced mainly by strombolian and surtseyan eruptions, with abundant cinder cones, tuff rings and subordinate lava flows. It is densely intruded by NE-, ESE- and WNW-trending swarms of dykes and sills that match the present volcanic vent systems and indicate that a triple rift system may have been an important feature of the El Golfo edifice. The lavas of the upper unit, in contrast, show a relatively small number of exposed feeder dykes, probably because they were mainly erupted from the summit of the volcano. These lavas show petrographic characteristics similar to (but becoming more evolved upwards than) those of the Tiñor volcano: olivine–augite basalts at the base, grading upward to microcrystalline lavas with scarce clinopyroxene phenocrysts. The more evolved trachytic rocks higher in the El Golfo edifice show abundant phenocrysts of alkali feldspar, aegirine–augite and opaque minerals in a trachytic matrix forming a network of

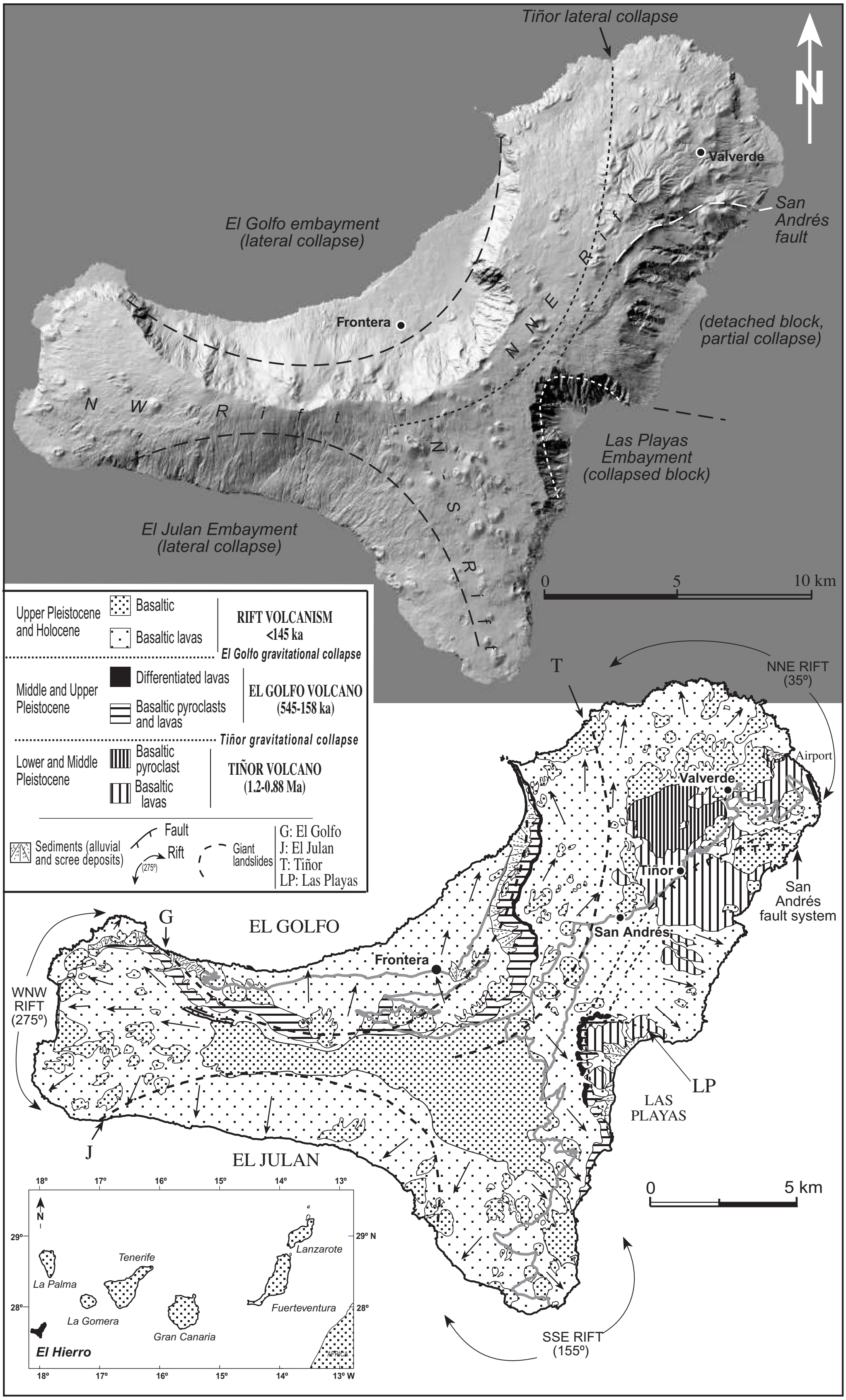

Fig. 18.11. Simplified geological map of El Hierro (from Carracedo *et al.* 1999*b*). Inset: shaded relief image of El Hierro with indication of the main geomorphological and tectonic features (image GRAFCAN).

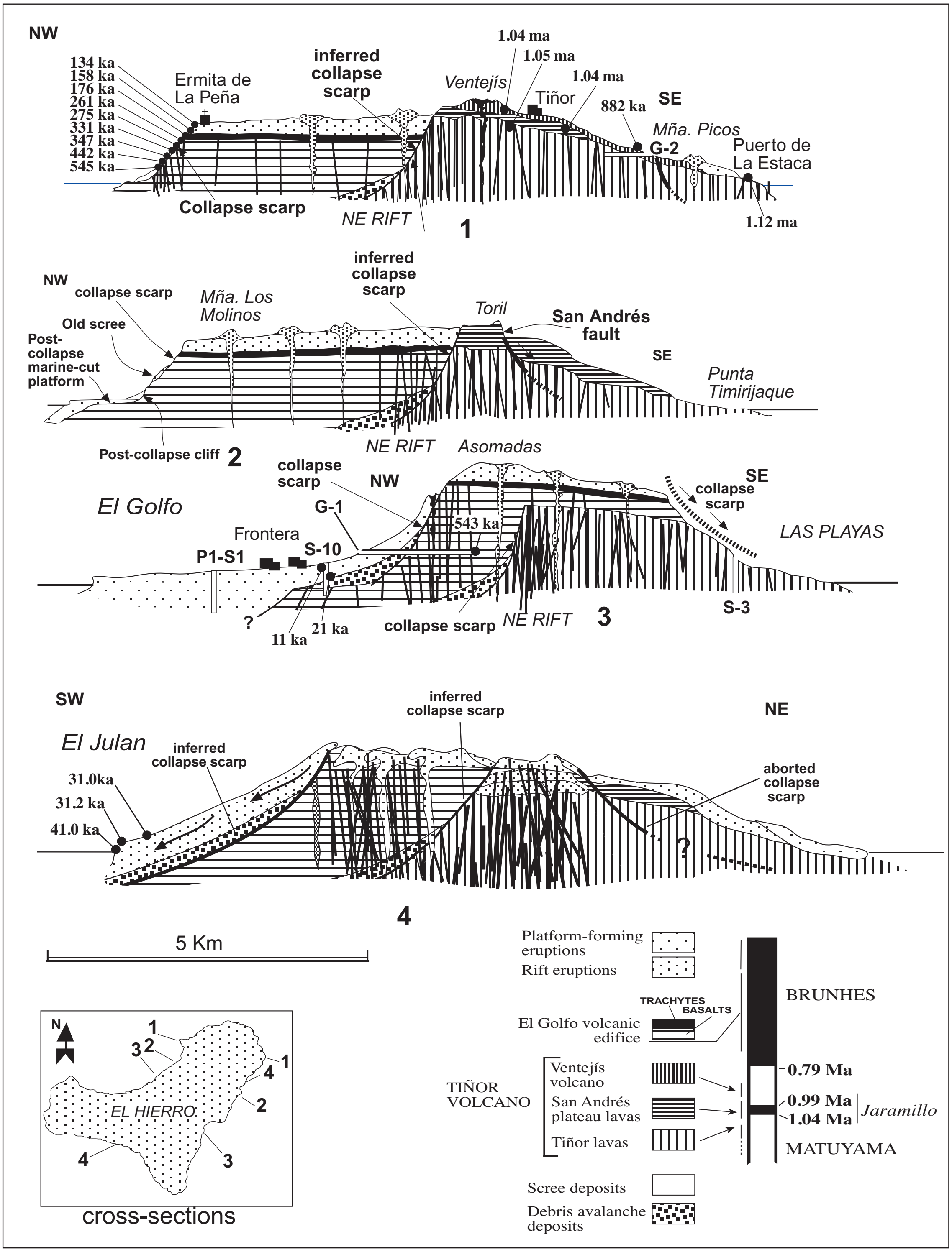

Fig. 18.12. Cross-sections (see lower inset map) illustrating the evolution of El Hierro by accretion of successive shield volcanoes (modified from Carracedo *et al.* 1999*b*).

feldspars and clinopyroxenes. The youngest rocks are formed by several differentiated lava flows (trachybasalts, trachytes) and block-and-ash deposits that probably correspond to the terminal stages of activity of the volcano, which was subsequently modified by lateral collapse. The El Julan collapse, on the SW flank of the volcano, is older than 158 ka, the age of the rift lavas that later partially fill this embayment. The SE flank underwent an aborted collapse between *c.* 261 and 176 ka BP, during which the San Andrés fault system (Fig. 18.11) was formed (Day *et al.* 1997). The youngest lateral collapse in El Hierro formed the northern El Golfo embayment and took place 15 ka BP ago according to Masson (1996) or between 21 and 134 ka BP according to Carracedo *et al.* (1999*b*).

Rift volcanism

The latest stage of growth of the island saw volcanism reorganized into a three-branched rift system, in which all the rifts were simultaneously active without the development of a central vent complex. Rift lavas show a slightly more alkaline and silica-undersaturated compositional trend, with alkaline picrobasalts, basanites and tephrites being the predominant petrographic types. The maximum age of these eruptions is constrained by the differentiated lavas topping El Golfo volcano (176 ka). K-Ar ages from 145 to 134 ka have been obtained at the top of El Golfo cliff (Fig. 18.12, 1). After this time the north flank of the island collapsed producing the present El Golfo embayment (Masson 1996; Carracedo *et al.* 1997, 1999*b*; Urgelés *et al.* 1997). Since this collapse there has been only moderate eruptive activity with lavas post-dating the last glacial maximum, filling the El Golfo embayment and forming coastal platforms. The last dated eruption (2.5 ka BP, Guillou *et al.* 1996) is from a small vent at the NE rift. However, an eruptive vent at the westernmost part of the island has been associated with the seismic crisis of 1793 (Hernández Pacheco 1982).

La Palma (JCC, ERB, FJPT)

The island of La Palma lies at the northwestern edge of the Canarian chain (Fig. 18.1), occupies a 706 km^2 triangular area, and forms a volcanic edifice that reaches a maximum elevation of 2426 m a.s.l. (Roque de Los Muchachos), resting on a 4000 m deep ocean floor. La Palma is formed mainly by two volcanoes: a circular, 25 km diameter shield volcano in the north (northern shield in Fig. 18.13 inset) and a north–south elongated, 20 km long Cumbre Vieja rift in the south (CV in Fig. 18.13 inset). The former, extinct for the last 400 ka, is deeply eroded, with a radial network of deep barrancos and a 6 km diameter erosional depression (Caldera de Taburiente) on its SW flank. The Cumbre Vieja volcano has been highly active since *c.* 120 ka BP, with six eruptions in the last 500 years, two of them in the twentieth century (1949 and 1971).

Early reports on the geology of La Palma were mainly focused on the spectacular Caldera de Taburiente and the submarine lavas cropping out in its interior, starting with the works of Buch (1825), Lyell (1865), Reiss (1861), Gagel (1908) and Sapper (1906). The submarine volcanic rocks have been studied by Middelmost (1970), Hernández Pacheco (1973), Staudigel (1981) and Staudigel & Schmincke (1984), among others. Geochronological, stratigraphic and structural studies have been carried out by Abdel-Monem *et al.* (1972), Ancochea *et al.* (1994), Guillou *et al.* (1998), Carracedo *et al.* (1999*a*) and Guillou *et al.* (2001), among others. Swath bathymetry and side-scan sonar studies of the marine flanks of La Palma (see Fig. 18.20B) have been published by Holcomb & Searle (1991), Weaver *et al.* (1992) and Urgelés *et al.* (1999).

The geological evolution of La Palma is characterized by the growth of three main volcanoes: a submarine volcano (seamount), the now extinct northern shield volcano, and the still active southern Cumbre Vieja rift (Fig. 18.13). Initial submarine magmatism produced alkali basaltic to trachytic pillow lavas and hyaloclastites, basaltic feeder dyke swarms, and gabbroic, trachytic and phonolitic intrusions. The entire complex has been affected by hydrothermal metamorphism, grading from zeolite to albite–epidote–hornfels facies, probably corresponding to a metamorphic gradient of *c.* 200–300°C/km (Staudigel & Schmincke 1984), and crops out on the floor and lower walls of the Caldera de Taburiente (CT in Fig. 18.13 inset).

The 20 km diameter submarine volcano has been dated at 4–3 Ma BP (Staudigel *et al.* 1986), and is separated from overlying subaerial volcanics by an angular unconformity. The angularity of this unconformity, and its present height of 1500 m, is due to uplift, tilting and erosion during the relatively long period required for the subaerial volcanism to become established, when soft, easily eroded pyroclastics were being produced. A 400–600 m thick sedimentary unit made of breccias, agglomerates and sediments lies between the submarine and subaerial volcanoes.

Following the emergence of La Palma above sea level around 1.77 Ma BP, two contiguous volcanic edifices were constructed. The earlier of these two subaerial volcanic centres forms the northern half of La Palma (Fig. 18.13) and provides an excellent illustration of a typical shield-stage Canarian volcano (Ancochea *et al.* 1994; Navarro & Coello 1994; Carracedo *et al.* 2001). The well-exposed sections provided by the Caldera de Taburiente walls allow the observation of the magmatic history from its initial stages, and over 100 water tunnels or 'galerías' permit the study and sampling of the internal structure of the volcanic edifice. Geomagnetic reversal data and precise radiometric dating (59 K-Ar and Ar^{40}/Ar^{39} ages) show that this shield is built of two subaerial volcanoes, Garafía and Taburiente volcanoes, developed from 1.77 to 0.4 Ma BP (Guillou *et al.* 2001).

The Garafía volcano developed discordantly over the seamount, increasing the island in areal extent and elevation, but subsequently became completely covered by the Taburiente volcano. Outcrops of the Garafía volcano (GV in Fig. 18.13 inset) are therefore rare and confined to erosional windows provided by the deep barrancos in the northern flank of the shield (Fig. 18.13). Continuous volcanic activity from 1.77 to 1.20 Ma BP built an edifice with steep flanks (slopes of 30–35°), 23 km in diameter and probably reaching 3000 m in height. Lava sequences around 400 m thick and comprising olivine–pyroxene and pyroxene–plagioclase basalts, with minor trachybasalts, constructed a 315 km^3 volcano with eruptive rates of 0.6 km^3/ka. Overgrowth and increasing instability triggered a giant lateral collapse towards the SW at *c.* 1.2 Ma BP (Figs 18.13 and 18.20). Breccias and sediments inside the Caldera de Taburiente (Fig. 18.13) may represent the avalanche deposits and post-collapse sediments associated with this first catastrophic tectonic event on La Palma.

The Taburiente volcano (TV in Fig. 18.13 inset) is the result of the resumption of the eruptive activity in the northern shield after the collapse of the Garafía volcano. From *c.* 1.1 to 0.4 Ma BP, a sequence of lavas over 1000 m thick completely covered the previous edifice and considerably enlarged the island in area and height. The first post-collapse lavas filled the collapse embayment and then spilled over the flanks of the Garafía volcano. Lavas flowing against the back scarp of the collapse formed a 400 m sequence of horizontal lavas which, with later differential erosion, subsequently formed a central plateau (CP in Fig. 18.13 inset) perched on the top of the shield. Volcanism then resumed at *c.* 0.8 Ma BP, changing

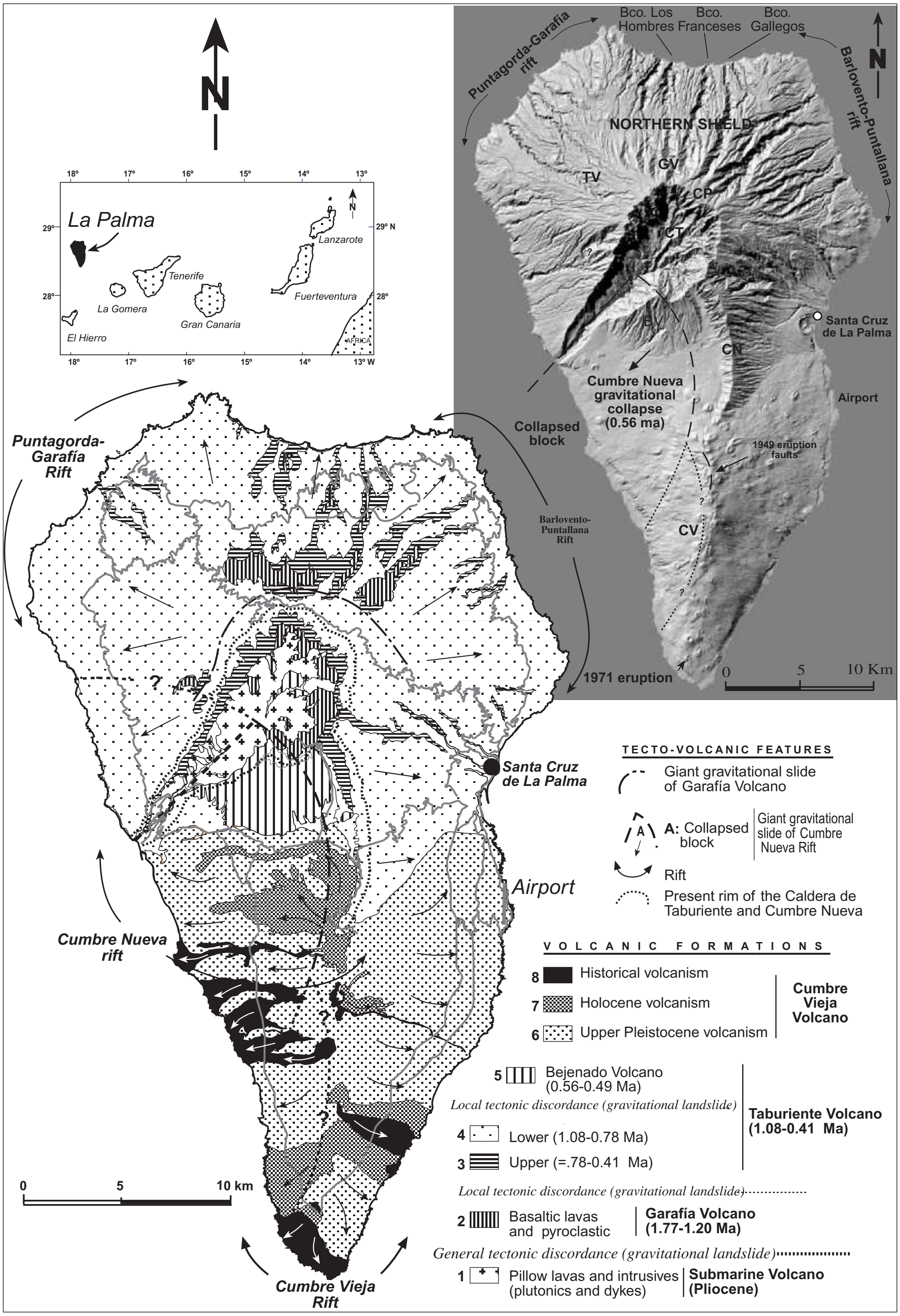

Fig. 18.13. Simplified geological map of La Palma (modified from Carracedo *et al.* 2001). Inset: shaded relief image of La Palma with indication of the main geomorphological and tectonic features (image GRAFCAN). Key: BV, Bejenado volcano; CP, central plateau (horizontal lavas filling the Garafía collapse embayment); CT, Caldera de Taburiente; Cumbre Nueva, Cumbre Nueva ridge; Cumbre Vieja, Cumbre Vieja volcano; GV, Garafía volcano; TV, Taburiente volcano.

from eruptions through dispersed vents in the shield to vents along well-defined rift zones (Figs 18.13 and 18.18A). In the later stages, a 3000 m high central volcano with differentiated lavas and explosive eruptions developed at the centre of the shield, together with rift-zone eruptions up until *c.* 0.4 Ma BP.

Volcanic activity may have already started to migrate southwards focusing predominantly in the southern Cumbre Nueva rift (CN in Fig. 18.13), which became progressively destabilized and collapsed towards the SW at *c.* 0.5 Ma BP. This collapse formed a prominent embayment (Fig. 18.13 inset) and initiated the 15 km long, 7 km wide, 2 km deep, *c.* 100 km^3 Caldera de Taburiente, first considered by Buch (1825) to be a typical 'uplifted crater', then an erosion caldera by Lyell (1865), and finally an avalanche caldera by Ancochea *et al.*

(1994). Studies by Carracedo *et al.* (1994, 1999*a*,*b*, 2001) and Paris & Carracedo (2001) concluded that this is an erosion caldera, initiated by a tectonic event, the Cumbre Nueva giant landslide. The main part of the depression developed by subsequent erosional retreat of the embayment walls. Linear incision rates in the caldera are often in excess of 1 m/ka and headward erosion rates exceed 3 m/ka, with the highest values at the boundaries of the collapse. Higher erosion rates in the western valley (Barranco de Las Angustias), coalescence of valleys and damming by lava flows of the Bejenado volcano led to the formation of a single circular depression. The Bejenado volcano (BV in Fig. 18.13 inset) developed inside this Cumbre Nueva collapse, representing a continuation of eruptive activity at the Cumbre Nueva rift, simultaneously with eruptions along other rifts of the shield. Remnants of the collapse embayment wall may still be present in slide blocks (Torevas) inside the caldera and underneath the Bejenado lavas.

The northern shield ceased volcanic activity at *c.* 0.4 Ma BP. However, apparently younger adventive vents on the Bejenado volcano may represent an eruptive interval between the northern shield and the Cumbre Vieja rift activity. Lavas of the Taburiente volcano show higher degrees of magmatic evolution, from primary basanites and basalts (olivine–pyroxene and amphibole) to differentiated rocks such as tephrites, phonolites and trachytes. The most differentiated lavas were erupted during the latest stages of the evolution of the volcano, and associated with more explosive eruptive mechanisms. Continuing southward migration of volcanism in La Palma ultimately resulted in the extinction of the northern shield *c.* 0.4 Ma BP and the formation in the last 130 ka of the Cumbre Vieja rift (CV in Fig. 18.13 inset). Eruptions from Cumbre Vieja occurred in two main pulses, one from *c.* 125 to 80 ka BP, and another from *c.* 20 ka BP to present (Carracedo *et al.* 1999*a*), building a 20 km long, 1949 m high ridge (Fig. 18.13). Reorganization of the rift from *c.* 7 ka BP apparently increased the instability of the volcano (Day *et al.* 1999), although the previous volcanoes of La Palma took considerably longer periods (0.5 and 0.8 Ma) to reach their instability threshold. The Cumbre Vieja volcano may collapse in the geological future or may evolve to a stable configuration.

Historical (<500 years) eruptions are located along the Cumbre Vieja rift (Figs 18.13 and 18.21), the most recent one (Teneguía volcano, 1971) being at its southernmost tip, with submarine vents extending the rift further to the south. An association of recent eruptions with phonolitic plugs is clearly evident, probably because these fractured plugs provide an easy pathway for the ascending magmas. Juvenile phonolites have been extruded in several of these recent eruptions, changing the characteristic effusive (strombolian) eruptions to more explosive mechanisms, with abundant block-and-ash deposits and magma mixing. Interaction with groundwater is frequent, and most of the recent eruptions show phreatomagmatic features (Kluegel *et al.* 1999).

Canarian geochronology, stratigraphy and evolution: an overview (JCC, FJPT)

Since the early K-Ar ages of Abdel-Monem *et al.* (1971, 1972), extensive geochronological studies have been carried out in the Canary Islands and more than 450 radiometric (K-Ar and $^{39}Ar/^{40}Ar$) ages from volcanic rocks of the different islands have been published. At least 105 of these ages, from volcanic lithologies in the islands of La Palma and El Hierro, have been obtained with stringent requirements: sampling from well-controlled stratigraphic sections, using only microcrystalline groundmass, replicated analyses, combined use of K-Ar and $^{39}Ar/^{40}Ar$ methods and systematic comparison of the palaeomagnetic polarities of the samples with the currently accepted geomagnetic reversal timescales (Guillou *et al.* 1996, 1998, 2001). A plot of the published radiometric ages from the Canaries (Fig. 18.14A) shows three groups of islands: (1) Lanzarote, Fuerteventura and Gran Canaria, with subaerial volcanism 14.5 Ma or older and two main stages of volcanic growth separated by long periods of inactivity (erosional gap in Fig. 18.14A); (2) La Gomera, with subaerial volcanism not older than 12 Ma and only the pre-erosional gap stage of growth; and (3) Tenerife, La Palma and El Hierro, with subaerial volcanism younger than *c.* 7.5 Ma and only the juvenile shield stage of growth.

An interruption in the volcanic activity (i.e. erosional gap) of individual islands is a common feature of hotspot oceanic island groups. This feature was used in the Hawaiian islands to separate two main volcanostratigraphic units: the shield stage and the post-erosional or rejuvenated stage (Clague & Dalrymple 1987; Walker 1990). The application to the Canaries of this distinction (Carracedo *et al.* 1998; Carracedo 1999) solves many of the problems raised by the original use of the term 'series' in the volcanostratigraphy of the islands (Fúster *et al.* 1968*a*,*b*,*c*,*d*), a term these days used for geological units formed during the same timespan and with synchronous boundaries. The use of terms such as 'old' and 'recent' series led to considerable confusion, since the 'old' series of La Palma or El Hierro is considerably younger than the 'recent' series of Fuerteventura, Lanzarote or Gran Canaria. This confusion is avoided when the concept of shield-stage and post-erosional or rejuvenation volcanism is applied to the Canaries, which can then be separated accordingly (Fig. 18.14B).

There are, however, some peculiarities in the stratigraphic relationship of the Canaries when compared to other oceanic groups such as the Hawaiian archipelago. In fact, a comparison of the two archipelagoes reveals significant differences: (a) in the age of the oldest islands (Fuerteventura is about four times older than Kauai); and (b) two islands in the Canarian archipelago are still in the shield stage, whereas in the Hawaiian islands only the active volcanoes on the island of Hawaii are at present in this stage of growth. Another important difference between the Canarian and the Hawaiian volcanoes is the extinction sequence (Stearns 1946; Clague & Dalrymple 1987). In the Hawaiian islands, the main stage of growth (typically lasting *c.* 1 Ma according to Langenheim & Clague 1987) of each island volcano is nearly completed before the next one emerges, whereas in the Canaries, three islands spanning at least 7.5 Ma are at present in their shield stage of growth.

In Fuerteventura, La Gomera and La Palma the products of shield volcanism rest upon variably deformed and uplifted sequences of submarine sediments, volcanic rocks (mainly pillow basalts), dyke swarms and plutonic intrusions which form the cores of these islands. These formations (termed the 'basal complex' by Bravo in 1964) are consistently separated from the subaerial volcanism by a major unconformity. Because of subsidence these pre-shield formations do not crop out in most of the volcanic ocean islands, except the Canaries and Cape Verde islands, notably on Maio (Stillman *et al.* 1982). Early interpretations linked these formations to uplifted

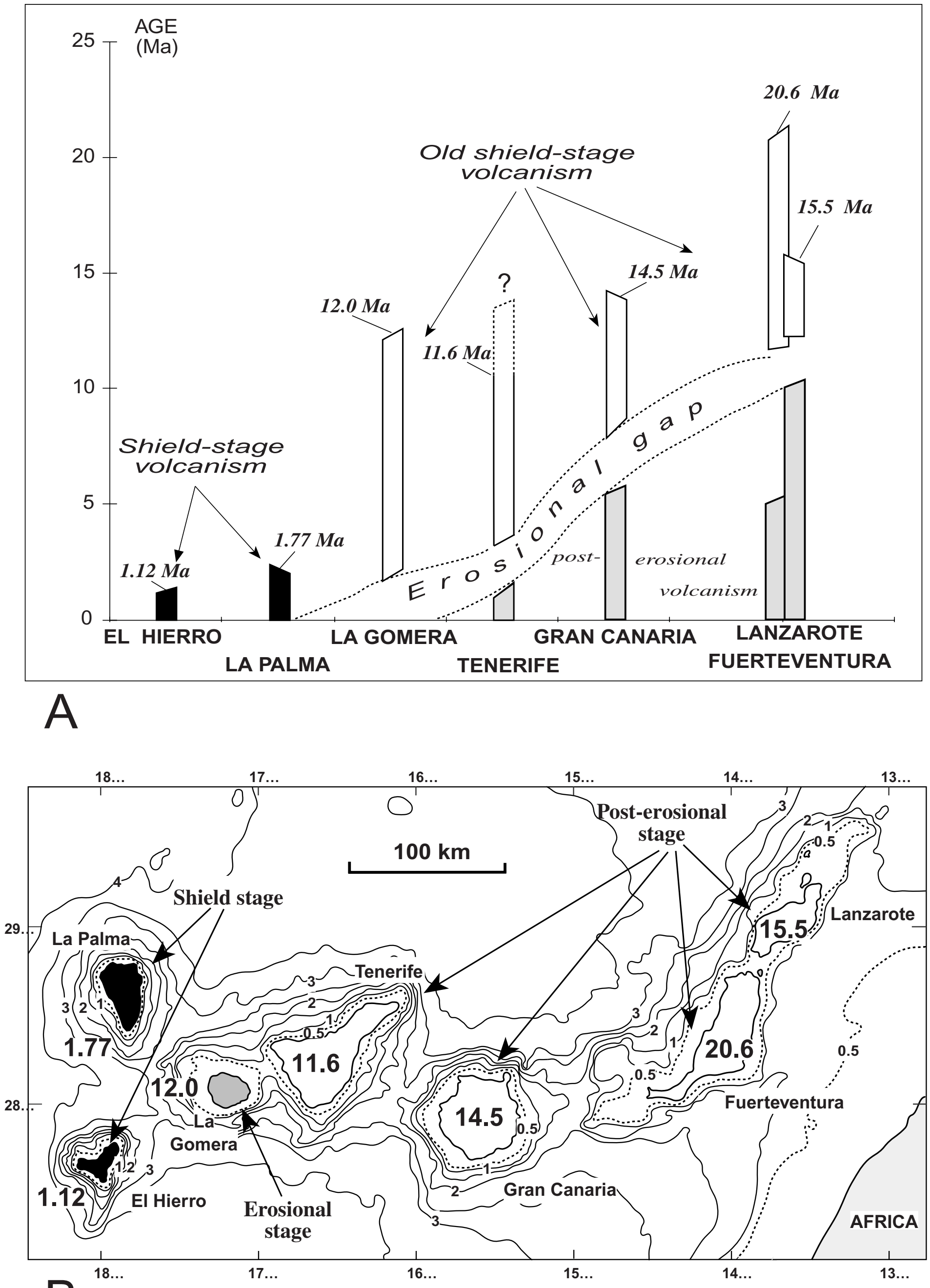

Fig. 18.14. (A) Published K-Ar ages from lavas of the Canary Islands. The presence of a gap in the eruptive activity allows the separation of two main stratigraphic units: shield-stage and post-erosional-stage volcanism. Canarian ages from Abdel-Monem *et al.* (1971, 1972), McDougall & Schmincke (1976), Carracedo (1979), Ancochea *et al.* (1990, 1994, 1996), Coello *et al.* (1992), Pérez Torrado *et al.* (1995), Guillou *et al.* (1996, 2001) X-axis not to scale. **(B)** Oldest ages of the subaerial volcanism of the Canary Islands. Sources of ages as in (A). (modified from Carracedo 1999).

blocks of 'oceanic basement' in the pre-plate tectonic sense (Hausen 1958; Fúster *et al.* 1968*b*,*c*). However, this interpretation proved to be inconsistent with the fact that the igneous rocks are younger than the oceanic sedimentary sequences (Robertson & Stillman 1979*a*,*b*). Studies in the Caldera de Taburiente in La Palma (Staudigel & Schmincke 1984) demonstrated that the 'basal complex' represents the seamount stage of the growth of these islands and, as anticipated in the Hawaiian group, in oceanic islands in general. Similar conclusions had been reached for the 'basal complex' of Fuerteventura (Stillman 1987).

Detailed geological mapping inside the Caldera de Taburiente (Carracedo *et al.* 2001) showed that several formations previously included in the 'basal complex' of La Palma could be assigned to younger subaerial stratigraphic units. When these units were excluded, the remaining formations conformed to the seamount described by Staudigel & Schmincke (1984). Similar circumstances are present in the geology of Fuerteventura and La Gomera, in which the rocks of the basal seamount are discordant with the subaerial volcanic rocks. We therefore propose that the term 'basal complex' be discarded and replaced by the general term 'seamount' or 'submarine volcanic edifice'. However, the latter terms have a genetic meaning, and would require, as in the subaerial volcanism, the definition of stratigraphic units and intrusive chronology.

Subsidence history

An important difference between the Canaries and other oceanic island groups (such as the Hawaiian islands) is the comparative absence of Canarian subsidence (Carracedo *et al.* 1998; Carracedo 1999). Individual islands in the Hawaiian group subside and eventually become seamounts. The amount, age and rate of subsidence have been derived from studies of submarine canyons, submerged coral reefs and coastal terraces of lava deltas (Moore & Fornari 1984; Moore 1987; Moore & Campbell 1987). The lack of significant vertical movements of the Canary Islands in their post-seamount stages becomes evident from the observation of the position of contemporary sea levels, in the form of marine abrasion platforms, littoral and beach sedimentary deposits, coastal volcanic deposits (hyaloclastite-based lava deltas and surtseyan tuff rings), and erosional palaeocliffs, all widespread in the Canary Islands. These features consistently occur close to present sea level, within the range of eustatic sea-level changes (Carracedo 1999). Marine abrasion platforms and marine and beach deposits interbedded with volcanic formations of different ages up to several million years old also appear consistently close to present-day sea level in Fuerteventura and Lanzarote (Meco & Stearns 1981; Carracedo & Rodríguez Badiola 1993). Surtseyan tuff rings appear at present sea level in the coastlands of the majority of the Canaries, as well as shallow pillow lavas of ages ranging from the Miocene to present day (Ibarrola *et al.* 1991; Carracedo & Rodríguez Badiola 1993).

Near-horizontal seismic reflectors observed in the volcanic apron of Gran Canaria (Funk & Schmincke 1998) show this island to have been stable at least since the late stages of shield building around 14 Ma ago. These reflectors reach the south flank of Tenerife, where they interbed with the volcanic aprons, providing evidence of the stability of this island during its entire volcanic history. It seems, therefore, that the islands of the Canarian archipelago, located close to the African continental margin, have been extremely stable, undergoing neither subsidence nor uplift after emergence.

The lack of post-emergence uplift, in contrast to the major uplift implied by the occurrence of the seamount formations, is also of interest since it implies that prior to emergence large intrusive complexes grew within the volcanoes. Conversely, post-emergence, endogenous growth of the islands was limited. This is consistent with the many geochemical and petrological data sets for subaerial volcanic suites in the Canary Islands, which indicate that these suites are for the most part fed by magma reservoirs in the underlying oceanic crust and/or oceanic lithosphere. Only the more evolved suites of rocks show some evidence for the presence of shallow magma reservoirs (Kluegel *et al.* 1999).

Different aspect ratios of the shield-stage and post-erosional-stage islands can be readily observed in the images of Figure 18.2. However, this is not a constructional feature since Fuerteventura and Gran Canaria may have reached similar or even greater elevations (Pérez Torrado *et al.* 1995; Stillman 1999), nor is it a consequence of subsidence, as discussed above, but of different stages of erosion. The Canaries (and the Cape Verde islands) apparently remain emergent for long periods of time, even exceeding 25 Ma, until completely mass wasted through gravitational collapses, relatively frequent in the juvenile stages of growth, and erosion.

Magma production rates and eruptive frequency

Another difference between Canarian volcanoes and the Hawaiian shields is that the latter involve much greater volumes and a higher frequency of eruptions (Walker 1990). The total volume of erupted magma and production rates are difficult or impossible to evaluate in the Canaries, especially for the western islands, in spite of the high quality and amount of age data available. The discontinuous character of volcanism, in which eruptive gaps, inherently difficult to date, may predominate over periods of activity, makes true evaluation of magma production rates unreliable unless large time intervals are compared. Most of the island edifices include non-volcanic sedimentary parts and the boundaries between islands are difficult to define. Furthermore, giant lateral collapses repeatedly removed large fractions of the mass of an island, especially during the shield-building stage, and redistributed them offshore over distances of hundreds or even thousands of kilometres. Several megaturbidites deposited in the Madeira abyssal plain within the past 1 Ma have been shown to have originated in the Canarian archipelago (Weaver *et al.* 1992). The rate of encroachment of volcaniclastic deposits onto the apron of Gran Canaria has been estimated to be of the order of >110 m/Ma in the shield stage, 55–110 m/Ma in the post-caldera, 22 m/Ma in the erosional gap stage and *c.* 66 m/Ma in the post-erosional stage (Schneider *et al.* 1998).

Despite these uncertainties, Schmincke (1982) computed some estimates of the total volume of the islands in the Canarian archipelago, including the products of submarine and subaerial volcanism, intrusions and sedimentary materials, but did not consider materials removed by mass wasting and by gravitational collapses. He obtained a range of estimated volumes of islands as follows: 5.3×10^3 km^3 for the smallest island of El Hierro to 23.8×10^3 km^3 for Gran Canaria and 30.6×10^3 km^3 for Fuerteventura. As pointed out by Schmincke (1982), the average total volume is remarkably similar to that of many shields in the Hawaiian islands (*c.* 20×10^3 km^3). This may imply that volumes of *c.* 20×10^3 km^3 are optimum values for the maximum growth of these oceanic islands, at least for the Canaries and the Hawaiian islands. However, actual values would require computing magma-production estimates using the combined volumes of the islands plus the offshore deposits.

Eruption rates during the subaerial shield-building stage, which are two to three orders of magnitude greater than those during the post-erosional stage, are therefore implied by considering a period of the order of tens of thousands of years. But even this may not be a sufficient averaging period because of switching activity between shield-stage islands on timescales of the order of hundreds of thousands of years. A further complication is the occurrence of episodes of relatively intense post-erosional volcanism such as that which produced the Roque Nublo volcano on Gran Canaria (Pérez Torrado *et al.* 1995).

The eruptive histories of the islands of El Hierro (Guillou *et al.* 1996) and La Palma (Guillou *et al.* 1998; Carracedo *et al.* 1999*a*,*b*; Guillou *et al.* 2001) are probably geochronologically the best constrained of any of the Canary Islands. The uncomplicated development of these islands, which are still in their juvenile stage of shield growth, together with the abundant and accurate K-Ar ages and magnetic stratigraphy allow the closest possible approach to the reconstruction of the entire emerged volcanic history of any of the Canaries. The present emerged volume of El Hierro, of *c.* 150 km^3, has been produced in the last 1.5 Ma, giving an apparent average magma production rate of 0.1 km^3/ka. However, rates increase significantly if we take into consideration the three known consecutive giant lateral collapses that affected the island, each clearly exceeding 100 km^3.

In La Palma, the northern Garafía and Taburiente volcanoes form a shield of *c.* 235 km^3 resting unconformably over the uplifted submarine edifice (Carracedo *et al.* 1999*a*,*b*; Guillou *et al.* 2001). In the southern half of the island, the Cumbre Vieja volcano has a volume of *c.* 130 km^3. Both have been constructed without important interruptions in their eruptive activity, the former between 1.7 and 0.4 Ma BP, and the latter between 0.15 Ma BP and the present. The eruptive rates for these volcanoes are, therefore, 0.18 and 0.86 km^3/ka, respectively. In comparison, average magma supply rates during the entire history of the island of Hawaii have been estimated to be of the order of 20 km^3/ka (Moore & Clague 1992). A similar evaluation of shield-stage magma production rates in the presently post-erosional islands is highly problematic. This is because it is impossible to evaluate the volume removed by lateral collapses (Stillman 1999). It is difficult to determine even the number of collapses in these deeply eroded islands, let alone the volumes of individual collapses.

Eruptive frequency and volume vary considerably, as observed for the historical eruptions over the last 500 years. The 1730–1736 eruption of Lanzarote is the largest to occur in the archipelago in this period, involving an eruptive volume as much as an order of magnitude larger than any other from historical eruptions in other Canarian islands (Carracedo *et al.* 1992). However, the previous eruption in Lanzarote may be that of the corona volcano, dated at 53 ka BP (Guillou, unpublished data). In the same period, as many as 100–1000 smaller eruptions may have taken place in the shield-building-stage islands of El Hierro, La Palma and Tenerife.

Petrology and geochemistry: an overview (EA)

The study of the rock composition of the Canary archipelago has been the subject of numerous works, especially by Fúster and co-workers and by Schmincke and collaborators (see Fúster 1975; Schmincke 1982). Over the last ten years, studies have focused on the isotopic composition of the magmas (e.g. Hoernle & Tilton 1991; Hoernle *et al.* 1991; Hoernle & Schmincke 1993; Marcantonio *et al.* 1995; Thirlwall *et al.* 1997; Thomas *et al.* 1999).

The plots of Fig. 18.15 show that the great majority of analyses of volcanic rocks fall in the alkaline, silica-undersaturated field of the total alkali versus silica (TAS) diagram, showing the typical variation trend of the alkali basalt kindred. There is a generally bimodal grouping into basalt–basanite and trachyte–phonolite compositions. A similar variation is found in the plutonic lithologies, pyroxenites, gabbros, alkali gabbros, ijolites and syenites and alkali syenites. On the other hand, with the exception of the Cape Verde islands, Fuerteventura is the only known oceanic island with exposed carbonatites (aegirinic, feldspathic and biotitic alvikites (Barrera *et al.* 1981), and sövites with sanidine, aegirine–augite, biotite, apatite, pyrochlore and magnetite (Demény *et al.* 1998)).

Around 1500 analyses of fresh volcanic rocks ($H_2O < 2\%$; $CO_2 < 1\%$) belonging to all these islands have been plotted on the TAS diagram (IUGS; Fig. 18.15), showing that most of them correspond to moderately alkaline (alkali basalts–trachytes) or highly alkaline (basanite–phonolite) rocks. Rocks of tholeiitic affinity (hypersthene-normative, but with no Ca-poor pyroxene) have been recognized only in the oldest units of Gran Canaria and in the most recent lavas of Lanzarote (Ibarrola 1969; Fúster 1975; Schmincke 1982; Carracedo *et al.* 1992). Ultra-alkaline rocks appear mainly in Gran Canaria (olivine melilitites and olivine nephelinites) and Fuerteventura (olivine nephelinites; Ancochea *et al.* 1996), corresponding in both these islands to the latest phases of their activity.

There are notable differences in the alkalinity and abundance of rock types between the islands (Fig. 18.15). Gran Canaria has rocks embracing all compositions from the most to the least alkaline, whereas in the other islands the alkalinity of lavas is more homogeneous. The most alkaline is La Palma, whereas Tenerife is less so (on the boundary between highly and moderately alkaline), and El Hierro, La Gomera, Fuerteventura and Lanzarote show the least overall alkalinity (all moderately alkaline).

The islands with the least felsic rocks (< 1%) are Lanzarote and Hierro, with even intermediate compositions being also very rare in Lanzarote. Fuerteventura is mainly formed of basaltic and trachybasaltic rocks, with *c.* 4% intermediate rocks and only around 1% trachytic rocks being present. La Palma and La Gomera have a somewhat higher overall abundance of felsic rocks (*c.* 3%), and in the central islands (Gran Canaria and Tenerife) evolved rocks are at their most important (> 10%).

Mg variation diagrams (Fig. 18.16) show similar trends in all islands, with decreasing MgO in basic rocks (basalts, basanites, nephelinites: MgO > 6%), both the CaO and the FeO contents decrease slightly, while Al_2O_3 rises considerably and TiO_2 has a tendency to increase slightly. These variations record the major role played by olivine and, to a lesser extent, clinopyroxene in the differentiation history of the magmas. Similarly, a change in the slope of the curves (decreasing in FeO, CaO and TiO_2) in rocks with <6% MgO (Fig. 18.16) records the influence of clinopyroxene crystallization control in the fractionation of tephrites, trachybasalts and basaltic trachyandesites. A further change of slope in the MgO–Al_2O_3 diagram shown by trachytic-phonolitic rocks (MgO < 1%) reflects the importance of plagioclase in the final stages of fractionation (Fig. 18.16). Trace element data confirm the previous trends, with the most incompatible elements (Fig. 18.17b) increasing from basalts to trachybasalts and basaltic trachyandesites (but not Ti). A change takes place in trachyandesites, with an important decrease in Ti and P, possibly due to Fe–Ti oxide and apatite fractionation. The decrease in Ti and P in trachytes and phonolites is accompanied by a Ba, Sr and Eu decrease, also observed in the rare earth element (REE) diagrams (Fig. 18.17c), in which trachybasalt and trachyandesites are enriched in REE content relative to the basalts, while the trachytic and phonolitic rocks show a negative Eu anomaly.

With respect to the primitive mantle, the rocks are enriched in a similar way in all the Canary Islands (Fig. 18.17a). A low amount of K and Rb suggests that a phase such as phlogopite or amphibole remains in the residuum at low degrees of partial melting (Hoernle & Schmincke 1993). The contents in trace element and radiogenic isotopes are characteristic of HIMU OIBs (Sun & McDonough 1989; Weaver 1991), although with some variations both between and within depending on the age of the various units (Hoernle *et al.* 1991; Thirlwall *et al.* 1997).

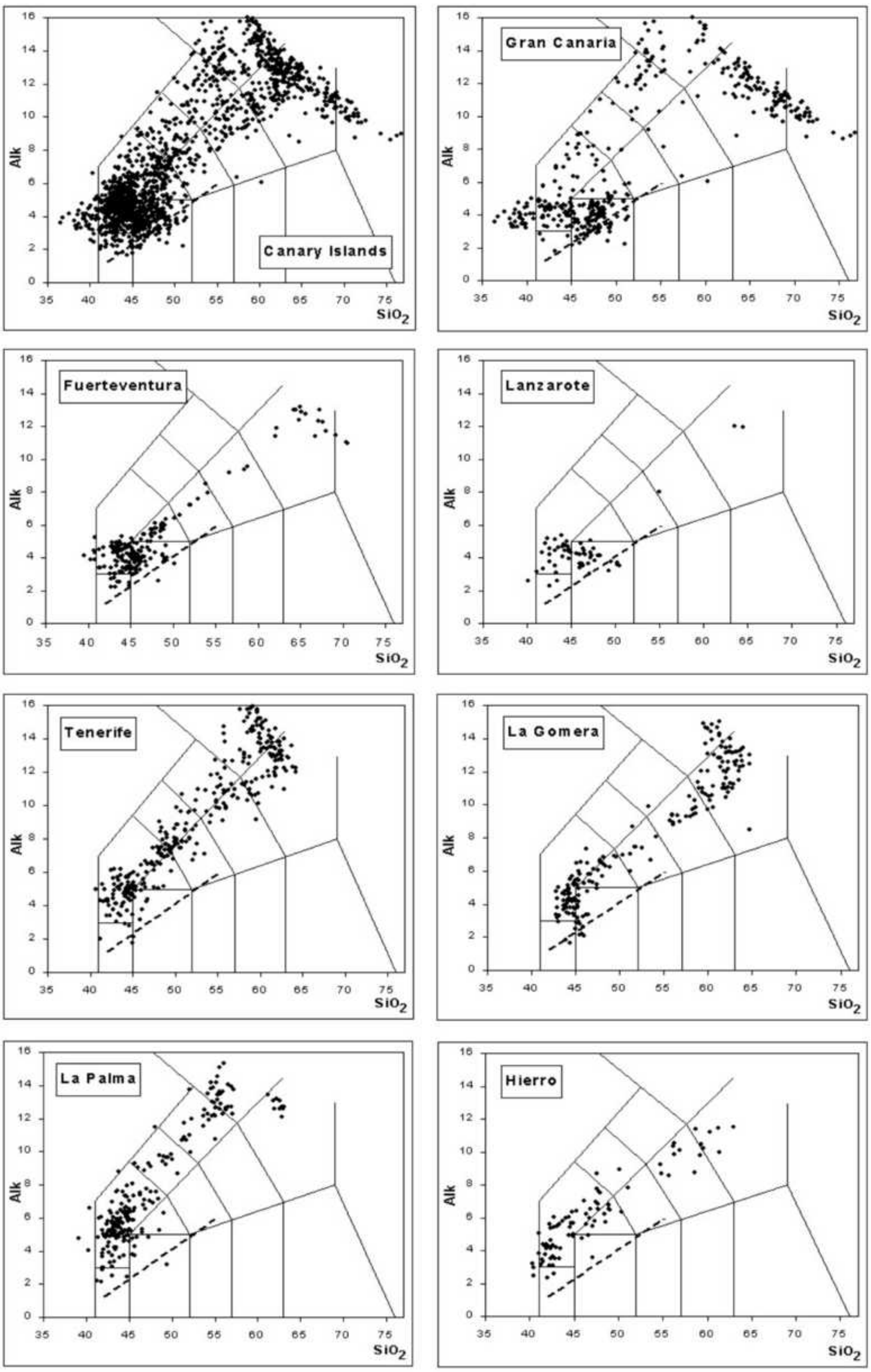

Fig. 18.15. Total alkali versus silica diagrams. Dashed line: MacDonald and Katsura (1964) alkali–tholeiite boundary. Data of the volcanic rocks ($H_2O < 2\%$; $CO_2 < 1\%$) from IGBA base data (Brändle & Nagy 1995) and Dpto. Petrología y Geoquímica (UCM) base data.

Canarian volcanic structure: an overview (JCC, FJPT)

The western and eastern Canaries show clear differences in structure and other important volcanic characteristics. The western islands display frequent, small-volume eruptions, high-aspect-ratio island edifices, well-defined, multibranched, long-lasting rifts and frequent massive flank collapses. Conversely, in the eastern islands volcanism is scarce and scattered, the islands have low aspect ratios, and rifts and giant landslides seem to be absent. Early interpretations related

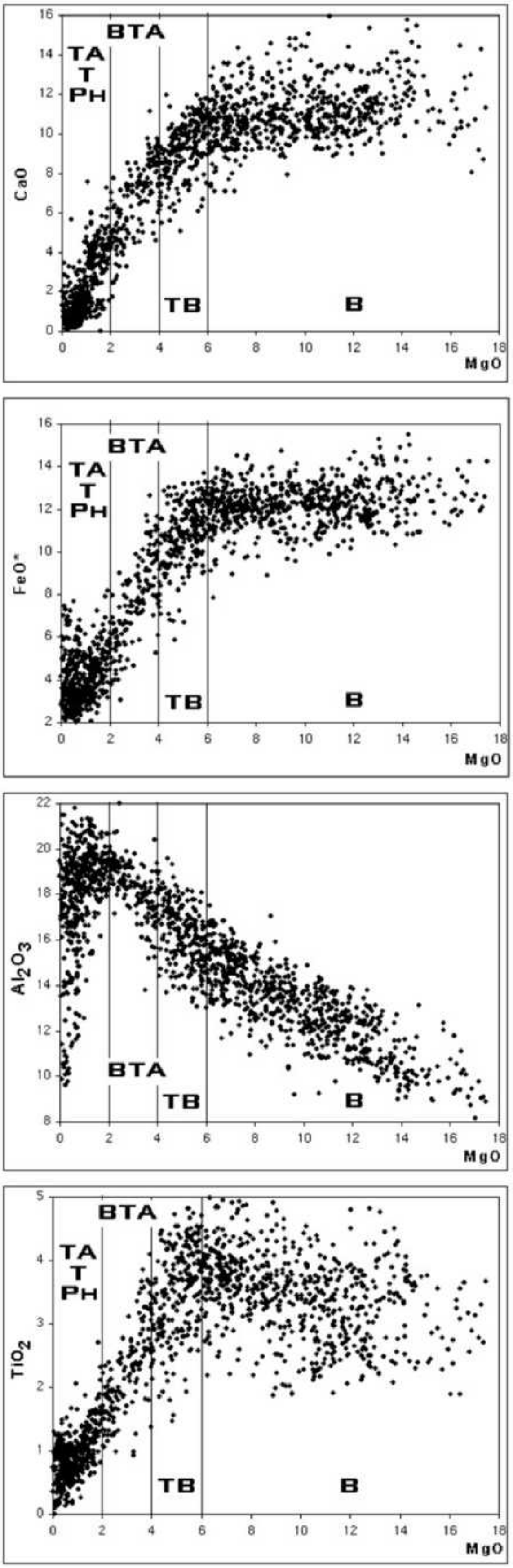

Fig. 18.16. Plots of MgO versus major oxides from Canary Islands volcanic rocks. Key: B, basaltic rocks; BTA, basaltic trachyandesites and phonotephrites; PH, phonolites;T, trachytes; TA, trachyandesites and tephriphonolites; TB, trachybasalts and tephrites.

these apparent contrasts to different geological and geophysical characteristics between the 'eastern Canaries', possibly underlain by continental crust, and the 'western Canaries', resting on oceanic crust (Rothe & Schmincke 1968; Dash & Bosshard 1969). However, subsequent studies have clearly shown the presence of Mesozoic oceanic crust beneath the entire Canarian archipelago (Schmincke *et al.* 1998; Steiner *et al.* 1998). These structural differences therefore instead reflect the different stages of evolution of the islands, with structural features readily observed in the young islands becoming removed by erosion in the older ones (Carracedo *et al.* 1998; Carracedo 1999).

The form, structure and landscape of the Canaries are characterized by four main features: (1) shield volcanoes, such as the Garafía-Taburiente in La Palma; (2) central stratovolcanoes, such as the Roque Nublo in Gran Canaria or the Teide volcano in Tenerife; (3) rift zones, locally known as 'dorsals', such as the Cumbre Vieja volcano in La Palma; and (4) collapse structures: vertical collapse calderas such as the Caldera de Tejeda in Gran Canaria, and gravitational collapse scarps and embayments such as the Caldera de Taburiente in La Palma or the Caldera de Las Cañadas in Tenerife.

Shield volcanoes

Several adjoining (or superimposed) shield volcanoes frequently form the bulk of the subaerial island edifices in the Canaries, as is also the case for the Hawaiian islands. Ancochea *et al.* (1996) showed that the island of Fuerteventura is composed of three shields (northern, central and southern), aligned in a NE–SW trend. This alignment continues in the adjacent island of Lanzarote with another two shield volcanoes (Los Ajaches and Famara volcanoes). The dimensions of these volcanoes vary from 30 to 47 km in diameter and they originally may have reached heights of up to 3 km above sea level (Stillman 1999). According to Ancochea *et al.* (1996), the duration of the volcanic activity in these shields varied from 3 to *c.* 7 Ma, although recent work shows that the ages of the volcanoes are considerably reduced when palaeomagnetic reversals and radiometric dating are applied with stringent requirements

A 50 km diameter, *c.* 2000 m high, shield volcano of *c.* 1000 km^3 developed in Gran Canaria apparently in a very short time (<500 ka; McDougall & Schmincke 1976; Bogaard *et al.* 1988; Bogaard & Schmincke 1998), building a predominantly effusive volcano, in which most of the dykes and vents are clustered in three convergent rift zones. The Gran Canaria shield-stage volcanism continued for more than 6 Ma, with a complex succession of processes including the formation of a collapse caldera and a cone-sheet swarm, and the emission of large volumes (*c.* 1000 km^3) of differentiated rocks (peralkalic rhyolites, trachytes and phonolites) in highly explosive eruptions.

In Tenerife, the ages of the two main shield volcanoes (Teno-Centro and Anaga) have been bracketed using geomagnetic reversals and radiometric dating (Abdel-Monem *et al.* 1972; Carracedo 1975, 1979). The Teno-Centro shield, 48 km in diameter, was constructed between *c.* 7.2 and 5.3 Ma BP,

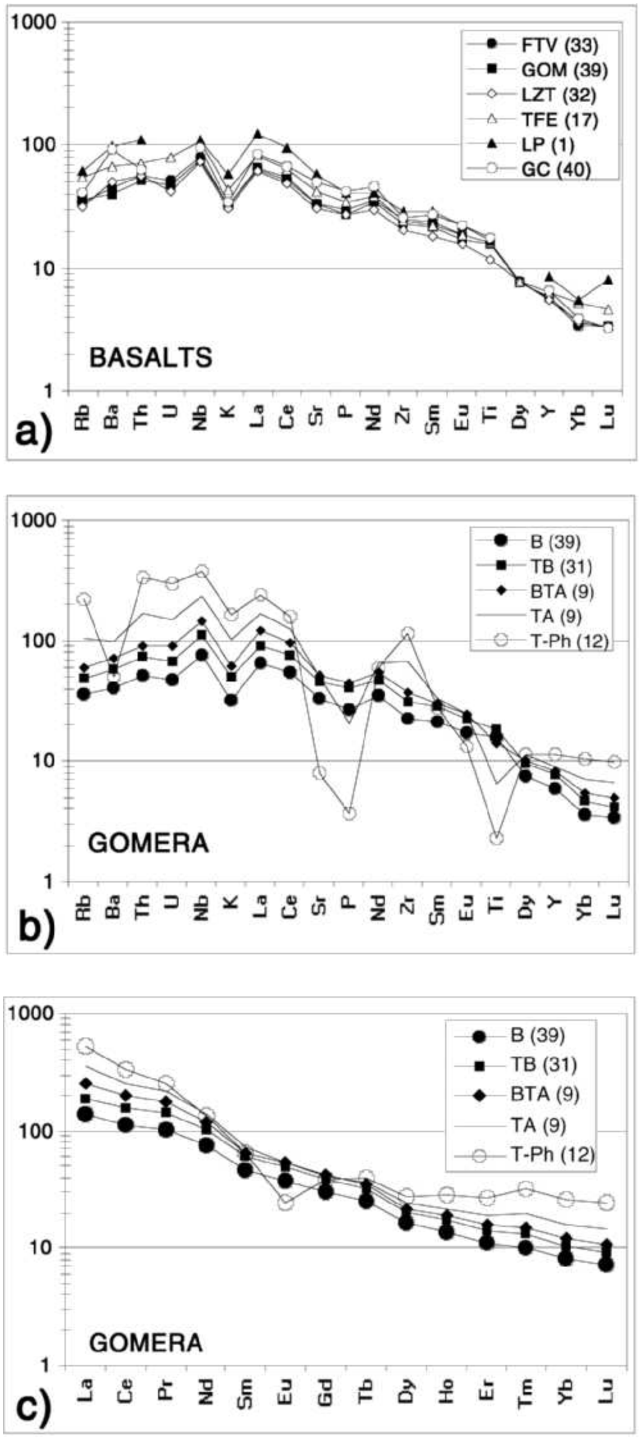

Fig. 18.17. Plots of incompatible trace elements normalized to primitive mantle (after Sun & McDonough 1989). **(a)** Basaltic rock averages (MgO > 6%). FTV, Fuerteventura; GC, Gran Canaria; GOM, Gomera; LP, La Palma; LZT, Lanzarote; TFE, Tenerife. **(b)** Rock averages from Gomera. **(c)** REE averages from Gomera normalized to chondrite (after Nakamura 1974). Abbreviations for rock types as in Figure 18.16. Gomera data from Dpto. Petrología y Geoquímica UCM (unpublished).

whereas the Anaga shield, 20 km in diameter, grew between 5.3 and 3.0 Ma BP. In a more recent work, Ancochea *et al.* (1990) defined three shields in Tenerife: Teno, Roque del Conde and Anaga, with time spans of 6.7–5.0, 8.5–6.4 and 6.5–4.5 Ma BP, respectively.

The Taburiente volcano in La Palma provides the best opportunity to define the characteristics (duration, volume, eruptive rates, magmatic evolution, etc.) of a well-preserved, entire shield. A similar volcanic history is observed in El Hierro, where near-continuous volcanic activity constructed the island from *c.* 1.12 Ma BP (Guillou *et al.* 1996). Eruptions produced a conspicuous three-branched, regular rift system and up to four consecutive lateral collapses (Carracedo *et al.* 1999b), one of them aborted (Day *et al.* 1997) with magmas evolving from basalts to trachybasalts and trachytes.

A duration of *c.* 1.5–2 Ma and an evolution from basaltic to mafic phonolites and trachytes seem to be characteristic of the western Canaries shield volcanoes. In the eastern Canaries, however, shield volume and duration may be considerably greater, as mentioned above, with diameters >30 km and growth time >3 Ma. Additional precise ages of these shield volcanoes are needed to obtain an improved understanding of their volcanic histories comparable to those of the shields in the western Canaries.

An important contrasting feature of the Canarian shields when compared with the Hawaiian shield volcanoes is that in the latter the volcanism in the most productive stage is typically tholeiitic, with minor volumes of alkali basalts and differentiated magmas. Silica-poor magmas (alkali basalts, basanites and nephelinites) predominate only in the rejuvenated volcanism of the post-erosional stages. In contrast, in the Canarian shields this difference is not evident and magmas of varied composition, from alkali basalts to phonolites and trachytes, are present in both stages.

Stratovolcanoes

Stratovolcanoes were built in different stages of the central Canaries: the Roque Nublo stratovolcano in Gran Canaria and the Teide–Pico Viejo volcano in Tenerife. The formation of stratovolcanoes requires the presence of shallow magmatic chambers and prolonged periods of repose to allow the differentiation of magmas. The Canarian stratovolcanoes have higher elevations and slopes (3718 m and 25–30% for Teide–Pico Viejo volcano in Tenerife, and 3000 m and 15–30% for Roque Nublo volcano in Gran Canaria) than the shields, but their basal diameters are considerably lower (*c.* 5.5 km and 5–10 km for the Teide–Pico Viejo and the Roque Nublo volcanoes respectively). Stratovolcanoes are, therefore, very unstable and gravitational collapses are more frequent than in shield volcanoes, although volumes are lower.

Parasitic vents and intrusions in the Canarian stratovolcanoes show a combination of radial and concentric trends (Martí & Mitjavila 1995; Pérez Torrado *et al.* 1995), whereas radial or three-armed rift zones predominate in shield volcanoes (Carracedo 1994, 1999). The compositions of lavas show variations from basic (alkali basalt–basanite) to silicic (trachyte–phonolite) compositions, the latter accounting for 20% of the total volume of the volcanoes. Crystal fractionation is the predominant process involved, although magma mixing and gaseous transfer processes have also been identified in the Teide–Pico Viejo and Roque Nublo volcanoes.

Eruptive mechanisms in these Canarian stratovolcanoes are significantly more explosive than in the shield volcanoes, with subplinian, plinian and vulcanian eruptions having been identified respectively in the Las Cañadas, Teide–Pico Viejo and Roque Nublo stratovolcanoes. Eruptive rates of Canarian stratovolcanoes have been calculated at 0.75 km^3/ka for Teide–Pico Viejo (Ancochea *et al.* 1990) and 0.1 km^3/ka for Roque Nublo (Pérez Torrado *et al.* 1995). A comparison of the volcanic history of the Roque Nublo stratovolcano and the Taburiente shield shows a much longer duration for the former (Fig. 18.18), involving *c.* 2 Ma, a period which includes 12 different magnetozones corresponding to the Gilbert–Gauss chrons. In contrast the Taburiente shield only lasted 1.4 Ma, and was erupted over six magnetozones in the Matuyama–Brunhes chrons. However, while volcanic activity occurred in the Taburiente shield during all the aforementioned magnetozones (Fig. 18.18A), there is evidence of volcanism in the Roque Nublo stratovolcano in only two of the 12 consecutive magnetozones (Fig. 18.18B). This clearly suggests long periods of continued volcanism in the Canarian shields, contrasting with short intervals of volcanism separated by long periods of repose in the stratovolcanoes.

Rift zones

Rift zones characterized by concentrations of vents and dense dyke swarms have been described in the majority of oceanic volcanic islands (Walker 1992). In the Canaries such zones typically form a tight cluster of eruptive vents aligned along narrow ridges, known locally as 'dorsals' (Fig. 18.19). These active volcanic structures show some characteristics of Hawaiian rifts (Tilling & Dvorak 1993), such as forceful injection of dykes, but they lack others, notably a direct connection with a shallow underlying summit magma chamber and caldera. Rift zones can be observed in exceptional detail in the Canaries, because the volcanoes are crossed at different depths and altitudes by many accessible water tunnels totalling >3000 km, excavated for mining groundwater (Carracedo 1994, 1996*a*,*b*). Their deep structure is characterized by numerous densely packed and closely parallel feeder dykes (Fig. 18.19A), with the swarm density increasing rapidly from the margins towards the axis of the rift zones and with depth. In outcrop, they form narrow zones of clustered eruptive vents, similarly increasing in density towards the axis of the rift zone and usually widening like a fan towards the lower flanks of the volcano (Fig.

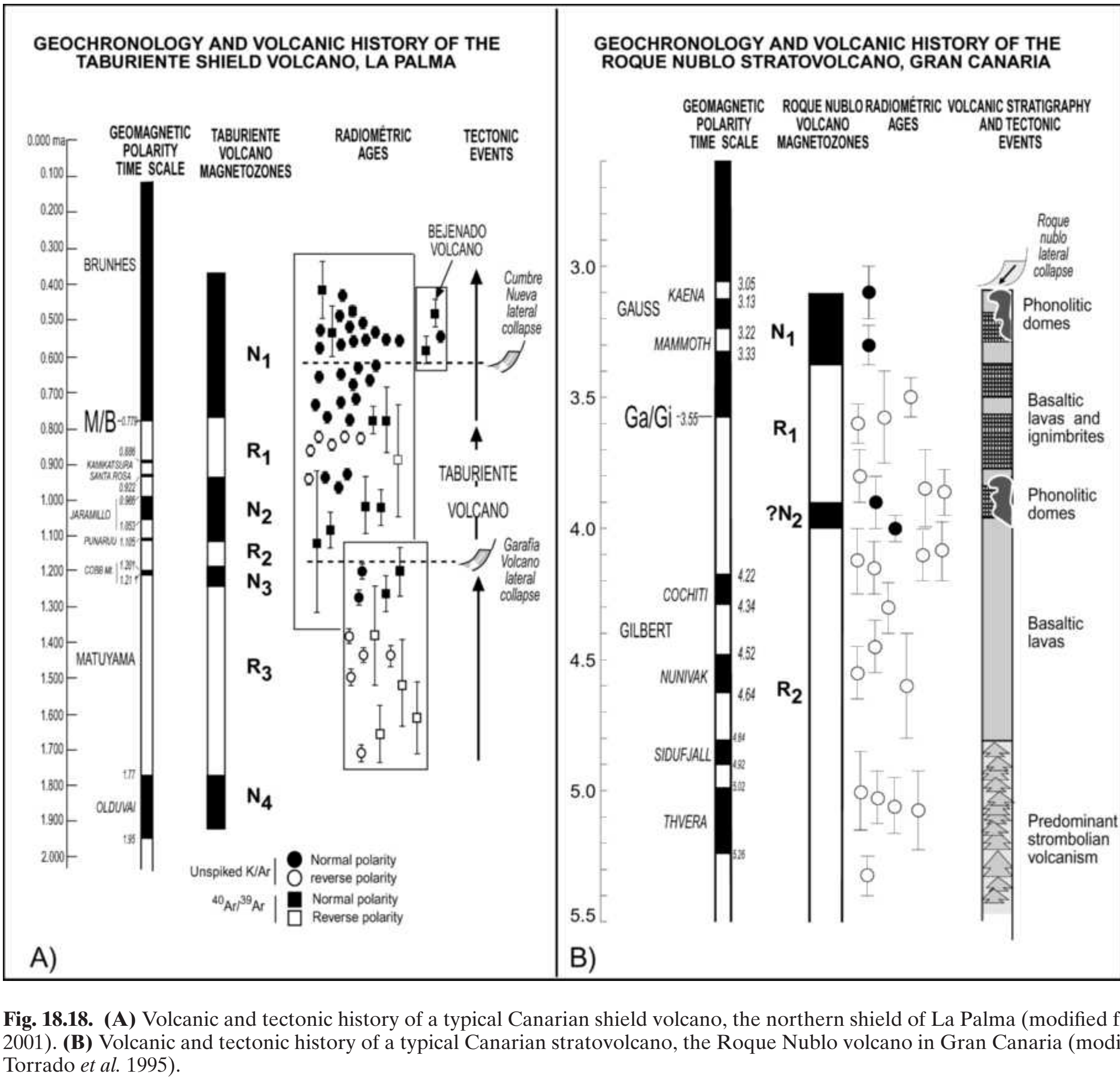

Fig. 18.18. (A) Volcanic and tectonic history of a typical Canarian shield volcano, the northern shield of La Palma (modified from Guillou *et al.* 2001). **(B)** Volcanic and tectonic history of a typical Canarian stratovolcano, the Roque Nublo volcano in Gran Canaria (modified from Pérez Torrado *et al.* 1995).

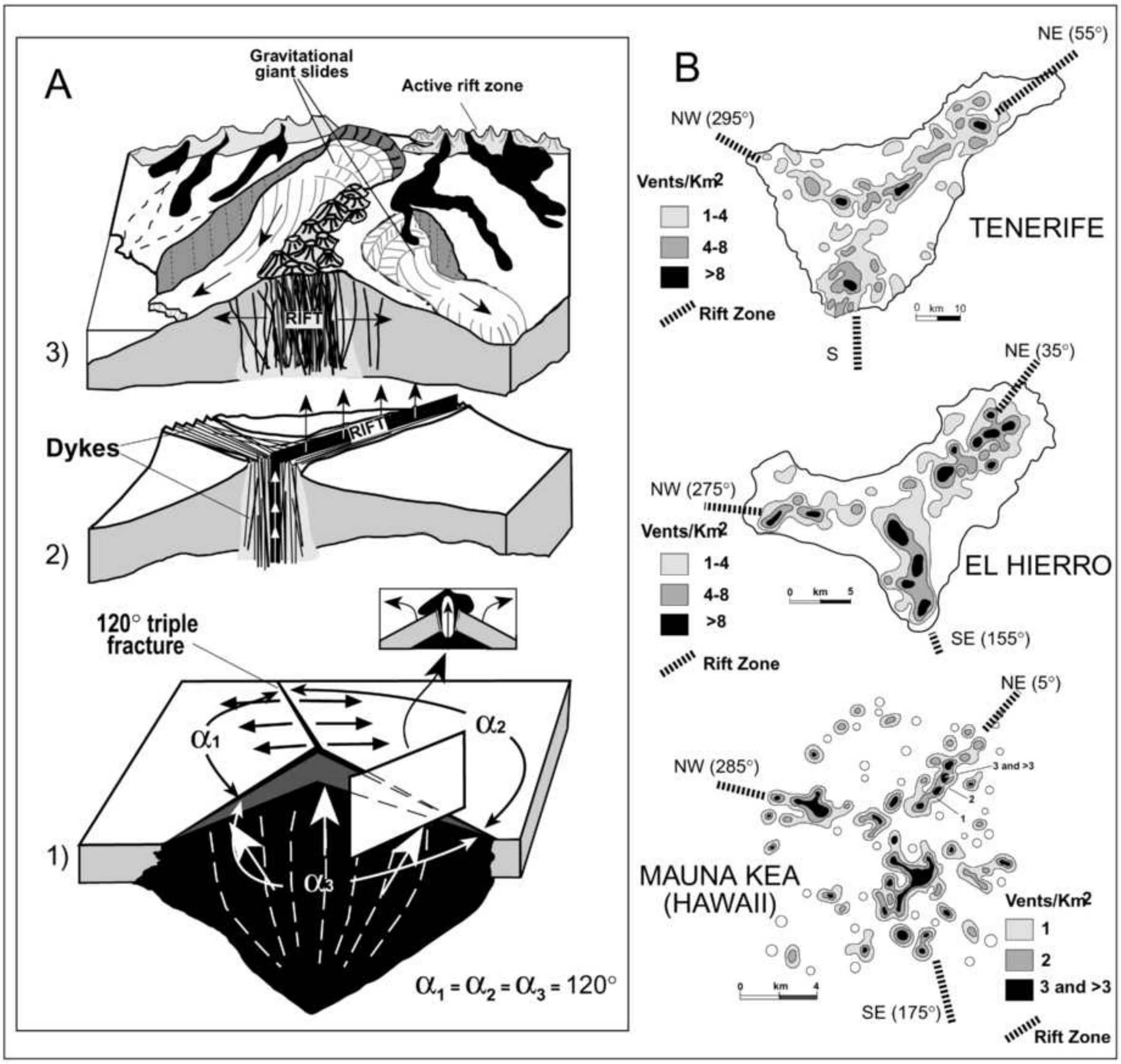

Fig. 18.19. **(A)** Model combining the formation of rifts and giant landslides in oceanic islands (modified from Carracedo 1994). **(B)** Distribution of eruptive vents in three-armed rifts in the Canaries. Mauna Kea vents from Porter (1972) (modified from Carracedo 1994).

18.19B). Walker (1992) proposed that rift zones in Hawaii contain about 65% of dykes and have a high bulk density (*c.* 2.8–2.9 mg/m^3), much greater than the host vesicular lavas (bulk density *c.* 2 mg/m^3). The surface rift zone geometry in the Canaries is characterized by three converging rifts, with angles of 120°, as in oceanic island volcanoes (Wentworth & MacDonald 1953; MacDonald 1972). Such a geometry is clearly observed in the distribution of Mauna Kea vents (Fig. 18.19B) mapped by Porter (1972), as seen in Tenerife and El Hierro in the Canary Islands (Carracedo 1994, 1999). However, the original regular geometry of rift zones is commonly lost during the evolution of the volcano, especially if buttressed among older volcanoes, as frequently happens in the very active initial shield-stage phases, such as seen in the island of Hawaii or in La Palma.

The presence of tripartite rift zones in the Canaries was first inferred by gravity measurements carried out by McFarlane & Ridley (1969). These authors observed a local Bouguer anomaly showing a three-pointed star shape coinciding with topographic ridges. They interpreted this feature as reflecting very high concentrations of dykes at depth, intruding along fissure zones, an interpretation that proved to be correct when the core of the rifts was observed in the 'galerías'. A simple model explaining the genesis of these features and their regular geometry was proposed by Carracedo (1994), in which three-armed rifts with angles of 120° resulted from least-effort fracturing due to magmatic intrusion (Fig. 18.19A, 1). Active rift zones in the Canaries are presently well developed only on those islands in the juvenile phases of shield building, with important recent eruptive activity (i.e. Tenerife, El Hierro and La Palma; Figs 18.19 and 18.21). This may reflect the observation of Vogt & Smoot (1984) that magmatism has to be frequent and intense for the rift zones to stay hot enough (thermal memory) to concentrate successive injections. Repeated magma injection swell deformation, and eventual rupture and resulting emplacement of blade-like dykes constitute a repetitive process that, by progressively increasing anisotropy, forces the new dykes to inject parallel to the main structure (Fig. 18.19A, 2; Carracedo 1994, 1996*a*,*b*). According to Walker (1992) there may be a limit to the growth of rift zones (65% in the Koolau dyke complex in Oahu, Hawaiian islands), with some mechanism restraining entry of magma into the complex after a certain intensity of intrusion and attendant widening are attained.

Collapse structures

Destructive collapse events, producing substantial changes in geomorphology and drainage systems, are the most conspicuous features of the Canarian landscape. Two genetic types can be defined: vertical collapse calderas, related to the abrupt emptying of magmatic chambers during highly explosive eruptions, and lateral collapse calderas and valleys associated with giant landslides. The best (and probably unique) observable vertical collapse caldera in the Canary Islands is the Caldera de Tejeda, in Gran Canaria, which formed some 14 Ma ago. This NW–SE orientated elliptical (20 × 17 km) caldera is located in what may have been the original summit of the basaltic shield of Gran Canaria. It has a fault system that separates at least three different blocks, each around 3 km wide, probably subsiding along listric faults into the caldera (Schmincke 1993). The vertical displacement of this fault system, with fault planes dipping 45° towards the caldera interior, has been estimated at *c.* 1000 m. Hydrothermal alteration of the intracaldera pyroclastic rocks through these fractures can be readily observed. From studies by Schmincke (1967), Hernán (1976) and Freundt & Schmincke (1992), the genesis of this collapse caldera may be related to the rapid growth of the basaltic shield of Gran Canaria, and the later abrupt emptying and collapse of a shallow rhyolitic magma chamber. The volume of ignimbrites associated with the caldera collapse has been calculated at *c.* 80 km^3, covering 400 km^2 of the SW flanks of the shield. After collapse, the caldera underwent a resurgence (Smith & Bailey 1968) due to the intrusion of a conical dyke swarm.

The lateral collapse events in the Canaries owe their origin to the fact that the volcanoes typically grow to a considerable height (Fig. 18.2), with gravitational stresses and dyke injections both tending to progressively increase the mechanical instability of the edifice, especially in the most active shield-stage phases of growth. The development of triple-branched rifts promotes this edifice overgrowth, with steep, unstable flanks and concentrations of dykes that destabilize the flanks through magma overpressure during emplacement (Swanson *et al.* 1976). These rifts induce mechanical and thermal pressurization of pore fluids, while extensional stresses develop in the axial zones of the rifts by dyke wedging, eventually exceeding the stability threshold and triggering massive flank failures. Slide blocks consistently form perpendicular to, or between, two branches of the rift system, with the remaining rift acting as a buttress (Carracedo 1994, 1996*a*,*b*). These collapses represent processes that oceanic volcanoes require in order to restore equilibrium and reduce altitudes that may restrain or suppress the internal plumbing system. They adequately explain the origin of morphological scarps, valleys and calderas (e.g. Caldera de Taburiente, Las Cañadas, valleys of La Orotava, Güímar) that were difficult to rationalize by erosive processes alone. The most recent known such event in the Canaries was the El Golfo landslide in El Hierro (8 in Fig. 18.20A) which took place around 120 ka ago (Carracedo *et al.* 1999*b*).

At least 11 of these giant collapses have so far been identified in the Canaries (Fig. 18.20A), both onshore (Ancochea *et al.* 1994, 1999; Carracedo 1994, 1996*a*,*b*, 1999; Carracedo *et al.* 1998; Stillman 1999) and offshore (Holcomb & Searle 1991; Weaver *et al.* 1992; Masson & Watts 1995; Watts & Masson 1995; Masson 1996; Urgelés *et al.* 1997, 1998, 1999). On occasions they can abort, as did the San Andrés flank collapse in El Hierro (Day *et al.* 1997), and probably one block may have detached from the Cumbre Nueva collapse in La Palma (see Fig. 18.13). Other examples may take place slowly, such as the gradual collapse of the southem flank of Kilauea volcano (Swanson *et al.* 1976; Smith *et al.* 1999). However, there is little doubt that these extremely low-probability events can produce disastrous effects. Two giant, and probably highly destructive, landslides occurred at *c.* 1.2 and 0.5 Ma BP, from the fast-developing Taburiente shield in La Palma (see Figs 18.18A and 18.20A, B). The average heights and time span required for the growth of these volcanoes to reach the instability threshold to trigger such a giant collapse seem to be *c.* 2500–3000 m and 0.5 Ma, respectively. In contrast, the Cumbre Vieja in southern La Palma is only 0.12–0.15 Ma old and <2000 m high.

Recent climatic changes (JM)

Post-Miocene erosion and sedimentation were influenced by palaeoclimate, so their analysis allows an attempt to reconstruct the climatic history of the Canary Islands over the last 6 Ma. Geoclimatic variations and related sea-level changes are especially well recorded by marine deposits, aeolian sands and calcretes in the eastern Canaries. Miocene–Pliocene marine deposits crop out in the islands of Lanzarote, Fuerteventura and Gran Canaria at +25 m to +55 m, +10 to +80 m and +70 m, respectively. They were originally interpreted as Quaternary in Fuerteventura and Lanzarote, and Miocene in Gran Canaria (Lyell 1865), but later palaeontological and geomorphological studies suggested a Messinian or early Pliocene age (Meco 1977; Meco *et al.* 1997). On Fuerteventura, pillow lavas with K-Ar ages of *c.* 5.8 and 5.0 Ma (Meco & Stearns 1981; Coello *et al.* 1992) overlie these marine deposits. In Lanzarote (Meco 1977) they are overlain by lavas with a K-Ar age of 6.7 Ma (Coello *et al.* 1992) and in Gran Canaria by pillowed lava with a K-Ar age of *c.* 4.37 and *c.* 4.25 Ma (Lietz & Schmincke 1975).

The presence of lower Pleistocene raised marine deposits in northwestern Gran Canaria at +85 m a.s.l. was first reported by Denizot (1934). Molluscs indicate an age on the Pliocene–Pleistocene boundary during a climatic interglacial similar to that of the present day. Middle Pleistocene marine terraces crop out along the northern coast of Gran Canaria at +35 m a.s.l., and are again interpreted as interglacial. Finally, upper Pleistocene marine deposits at +12 m a.s.l. . on the NE coast of Gran Canaria, first reported by Lyell (1865) and Rothpletz & Simonelli (1890), are similar in age and faunal composition to the last interglacial marine deposits of the south coast of Fuerteventura and Lanzarote at *c.* +5 m a.s.l. (Meco *et al.* 1997).

During post-Messinian regression, coinciding with the establishment of the cold Canary current, Pliocene bioclastic aeolian sands entirely covered the islands of Fuerteventura, Lanzarote and the NE of Gran Canaria to elevations of

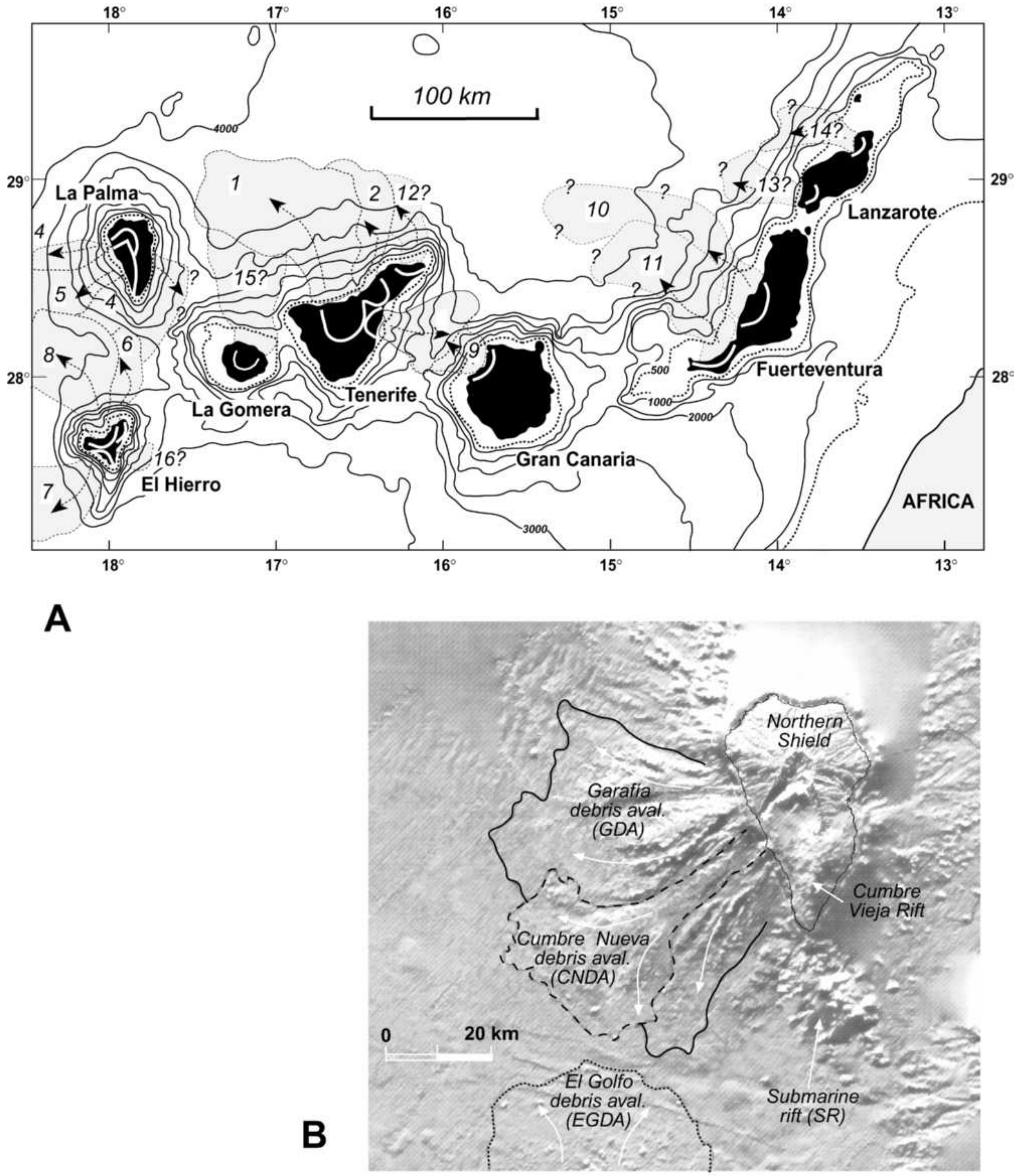

Fig. 18.20. **(A)** Documented and inferred giant landslides in the Canaries (modified from Urgelés *et al.* 1995, 1997; Carracedo *et al.* 1998). **(B)** Shaded relief images of bathymetric and topographic data of La Palma showing debris avalanches and related giant landslides (modified from Urgelés *et al.* 1999).

c. 400 m. These biocalcarenites include terrestrial gastropods, tortoises, eggshells of large birds and alluvial horizons with angular volcanic clasts, fragmented Messinian marine fossils and vertical roots. After the installation of a drainage system, middle Pleistocene–lower Holocene loessic palaeosols interbedded between aeolian sands record humid spells in a generally arid regime (Meco & Petit-Maire 1986; Meco *et al.* 1997).

Calcrete on Gran Canaria was first described by Buch (1825), and on Lanzarote and Fuerteventura by Hartung

(1857). The calcrete is frequently several metres thick and was formed on top of the aeolian sands by dissolution of carbonate fragments of the marine fossil flora and fauna. This process took place during the climatic change from wet and warm to arid and cold at the beginning of the first Quaternary glaciation. Wind reworking of the underlying sands through erosion windows in the calcrete formed the middle Pleistocene to present dunes and beaches of the Fuerteventura and Lanzarote islands, although Pleistocene dunes are also covered by very thin calcretes.

Holocene volcanism and modern hazards

Holocene volcanic eruptions have occurred across the entire Canarian group with the exception of La Gomera. However, the intensity of Holocene volcanism is not constant throughout the archipelago, with the shield-stage islands having eruptive frequencies and volumes at least 10–100 times greater than those for the post-erosional islands (Fig. 18.21). Ten to 100 eruptions have been identified in La Palma (Carracedo *et al.* 1999*a*; Guillou *et al.* 2001), El Hierro (Guillou *et al.* 1996) and Tenerife (Fúster *et al.* 1968*a*), whereas less than ten eruptions took place in Gran Canaria (Fúster *et al.* 1968*d*; ITGE 1992), Fuerteventura (Fúster *et al.* 1968*b*) and Lanzarote (Fúster *et al.* 1968*c*; Carracedo *et al.* 1992). The occurrence of a relatively very long and voluminous eruption in Lanzarote (1730–1736) has prompted the erroneous idea of similar levels of volcanic activity along the entire Canarian chain. However, this idea is inconsistent with the fact that, prior to the 1730 eruption, the only eruptions of note on Lanzarote were those of the Corona volcano (53 ka) and Los Helechos volcanic group (72 ka).

Geochronological control for the late Pleistocene–Holocene volcanism has still been poorly addressed, with the exception of the island of La Palma. Volcanism preserving youthful morphology has generally been grouped and mapped as 'Recent', 'Quaternary' or 'Basaltic Series III and IV' (Fúster *et al.* 1968*a*,*b*,*c*,*d*). However, differences in age are significant, probably due to important variations in erosion rates in relation to changes in climatic conditions and/or geographic location. The use of precise radiometric dating and sea-level changes related to glaciations allowed the definition of several stratigraphic subunits in the upper Pleistocene–Holocene successions of La Palma (Carracedo *et al.* 1999*a*), defining more precisely the stratigraphy and mapping of the volcanic activity. The Cumbre Vieja volcano, previously considered older than 620 ka (Abdel-Monem *et al.* 1972) or 540 ka (Ancochea *et al.* 1994), has been constrained to *c.* 120 ka (Guillou *et al.* 1998; Carracedo *et al.* 1999*b*).

Historical eruptions during the last 500 years that occurred in the Canary Islands are shown in Figure 18.21. These eruptions have happened when the islands had a low population or, later, in places with very few inhabitants. Only the 1706 eruption of Garachico on Tenerife, the anomalously long 1730 eruption of Lanzarote and the most recent eruptions of La Palma (1949 and 1971) posed any significant threat to the people and the economy of the region. Notwithstanding, the spectacular increase in the Canary Islands' population (1.8

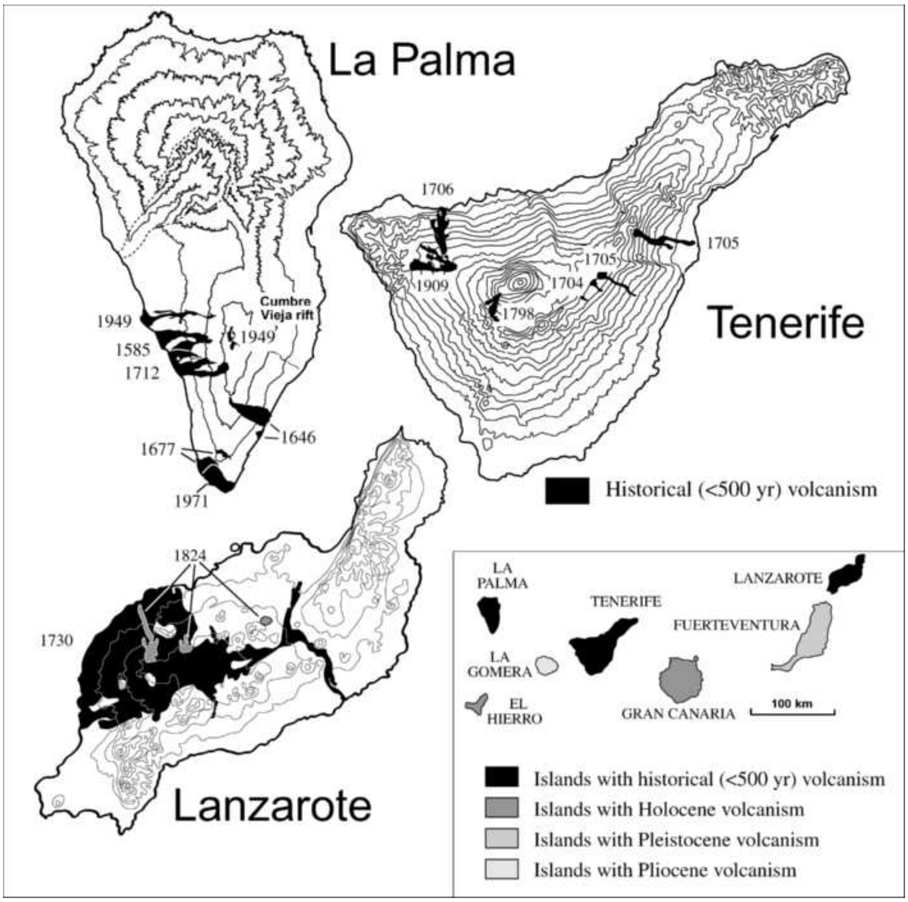

Fig. 18.21. Recent (inset) and historic (<500 years) volcanism in the Canary Islands (modified from Carracedo 1994). Note that most of the historic flows reach the coast in La Palma (juvenile island) and not in Tenerife (post-erosive).

million inhabitants and 10–11 million visitors annually) has accordingly increased the risks.

The first eyewitness account of a volcanic eruption in the Canaries is that of Torriani on the 1585 Jedey eruption in La Palma (Torriani 1592). Important errors in the identification of which vents were active and which lava flows were produced in the different eruptive events are common in the geological literature, even in cases of historical volcanism with contemporary descriptions, such as the 1585 and 1677 eruptions of La Palma (Santiago 1960; Carracedo *et al.* 1996). Modern eruptions on La Palma and Tenerife are clearly related to the rift zones, whereas in Lanzarote they are located in the central, NE–SW tectonovolcanic feature. Historical eruptions in the Canaries are characteristically basaltic fissure eruptions, with the Jedey or Tahuya eruption (1585, La Palma) producing both basalts and juvenile phonolites, and the 1730 eruption of Lanzarote producing tholeiites (Carracedo & Rodríguez Badiola 1991; Carracedo *et al.* 1992). Eruption duration varies from eight days to more than six years, but is typically one to three months. Volumes range from 0.2×10^6 to *c.* 700×10^6 m^3, most commonly being 10×10^6 to 40×10^6 m^3. The area covered by lavas during an eruption varies from 0.2×10^6 to 150×10^6 m^2, with values generally between 3×10^6 and 10×10^6 m^2. All historical eruptions showed seismic precursors, from less than one year to only a few hours prior to eruption onset.

Although El Hierro is geologically the youngest island of the Canaries, most of the recent eruptions have occurred in La Palma where at least ten eruptive events have occurred over the last 2500 years, compared with only one eruption in El Hierro in the same period (Mña. Chamuscada, ^{14}C age of 2500 ± 70 years BP; Guillou *et al.* 1996). Recent felsic, explosive volcanism is limited to the Teide–Pico Viejo volcanic complex in Tenerife. However, according to Barberi (1989), the occurrence of several basaltic eruptions adventive to this complex suggests that its magmatic chamber is very reduced in size or inactive, an observation in agreement with the lack of explosive eruptions parallel to these basaltic events.

The most likely potential volcanic hazards in modern times are related to basaltic fissure eruptions in the shield-stage islands, particularly in the Cumbre Vieja volcano of southern La Palma, where the remote possibility of a catastrophic collapse (with resulting tsunami destruction) also exists. However, the absence of any detectable displacement of the unstable western flank (Moss *et al.* 1999) and the lack of seismicity seem to exclude any immediate hazard. From available information for older giant collapse in the Canaries, it may take tens, or even hundreds of thousands of years for the Cumbre Vieja volcano to become critically unstable, assuming that the volcano does not evolve to stable configurations in the geological future. Furthermore, whereas the Garafía and Cumbre Nueva volcanoes developed for 570 and 640 ka respectively before collapsing (Carracedo *et al.* 1999*a*,*b*; Guillou *et al.* 2001), the age of the Cumbre Vieja volcano is only *c.* 120 ka.

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