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THE ITALIAN QUATERNARY VOLCANISM

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ABSTRACT: The peninsular and insular Italy are punctuated by Quaternary volcanoes and their rocks constitute an important aliquot of the Italian Quaternary sedimentary successions. Also away from volcanoes themselves, volcanic ash layers are a common and frequent feature of the Quaternary records, which provide us with potential relevant stratigraphic and chronological markers at service of a wide array of the Quaternary science issues. In this paper, a broad representation of the Italian volcanological community has joined to provide an updated comprehensive state of art of the Italian Quaternary volcanism. The eruptive history, style and dynamics and, in some cases, the hazard assessment of about thirty Quaternary volcanoes, from the northernmost Mt. Amiata, in Tuscany, to the southernmost Pantelleria and Linosa, in Sicily Channel, are here reviewed in the light of the substantial improving of the methodological approaches and the overall knowledge achieved in the last decades in the volcanological field study. We hope that the present review can represent a useful and agile document summarising the knowledege on the Italian volcanism at the service of the Quaternary community operating in central Mediterranean area.

Keywords: Italian volcanisms, stratigraphy, geochronology, eruptive dynamics, distal tephra.

1. INTRODUCTION (S.C., B.G., P.P.)

Volcanic products undoubtedly represent an important part of the Quaternary deposits in Italy. For this reason, we have decided to take the opportunity offered by the INQUA conference to be held in Rome in 2023 to present a review of the state of the art and knowledge of the Italian Quaternary volcanism. A few remarks should be made to explain why we felt the need to do this.

The activity of the Italian Quaternary volcanoes during the last century has been the subject of intense investigations, although at first the main focus was on mineralogical-petrographical-magmatological aspects other than on the strictly volcanological ones. It was probably in the decade 1980-1990, with the publication of the results of the "*Progetto Finalizzato Geodinamica*" of the Italian National Research Council (CNR), that monographic works were published on some volcanic areas, such as the Neapolitan volcanoes or Etna, which, in addition to the petrographic features, proposed a systematic study of the stratigraphic aspects of the products, mapping their distribution and thus providing a reconstruction of their eruptive history (e.g., Santacroce, 1987 for Somma-Vesuvio, Rosi & Sbrana, 1987 for Campi Flegrei, Vezzoli, 1988 for Ischia, Romano et al., 1979 for Etna, etc.).

Since then, and especially in the last two decades, this reconstruction has been progressively refined due to the improved knowledge of volcanic history and stratigraphy, in particular thanks to the Cartografia geologica e geotematica (CARG) project (e.g. Groppelli & Vireck-Goette, 2010), dedicated to updating the geologi-

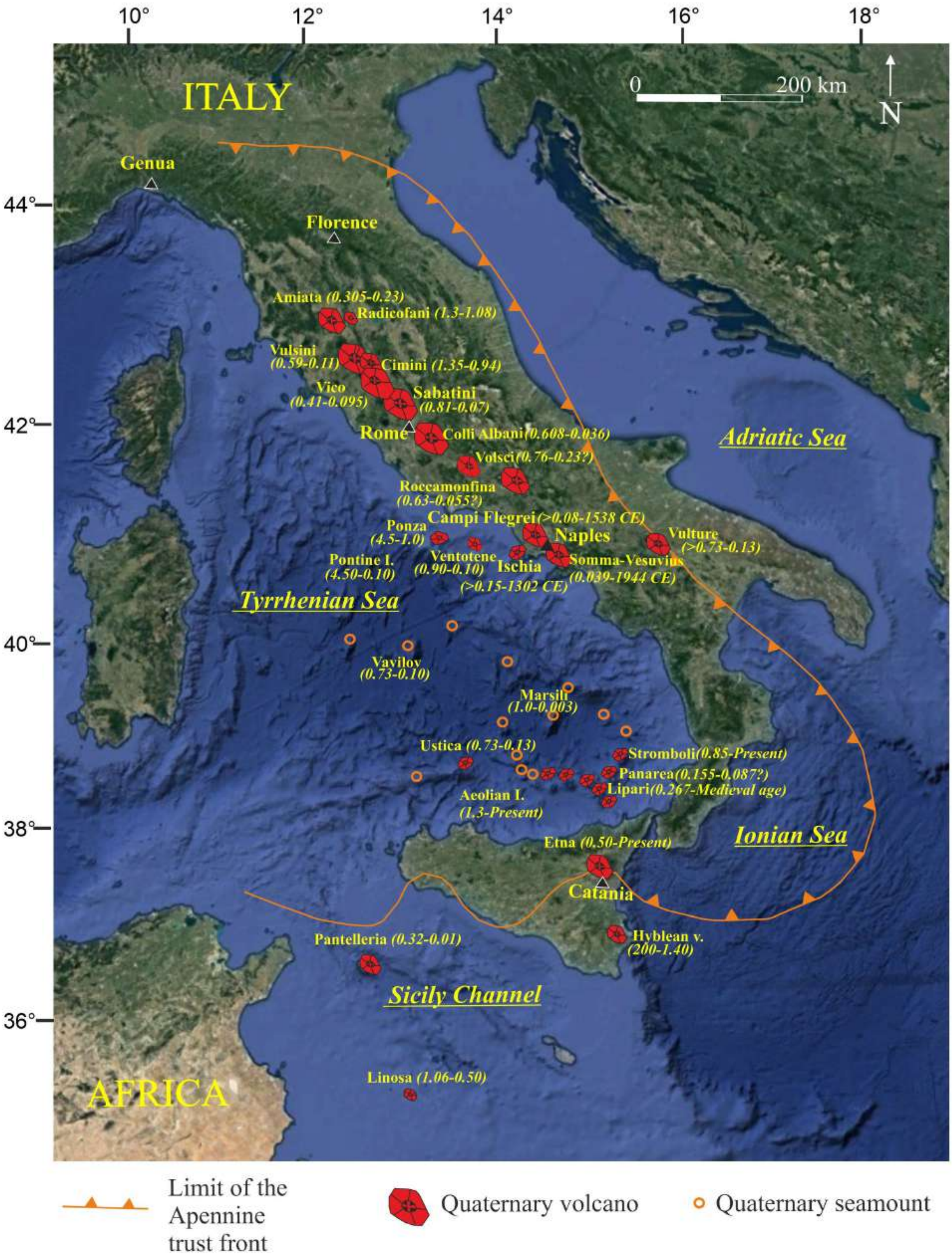


Fig. 1 - Location of the Italian Quaternary volcanoes with indication of the time interval of activity (Ma, unless differently indicated).

cal mapping of the Italian territory. A multidisciplinary approach (stratigraphic, geomorphological, geochemical) allowed the reconstruction of the time-space history of most of the Italian Quaternary volcanic areas, which has been furthered by the increased ability to obtain reliable $^{40}Ar/^{39}Ar$ ages on both pyroclastic deposits and lavas. In fact, the significant technological developments of noble gas mass spectrometers over the last decade (Mixon et al., 2022), have improved the effectiveness of the method and the possibility of getting accurate and high-precision $^{40}Ar/^{39}Ar$ dating allowing a detailed and robust reconstruction of the eruptive history. This has been made possible and facilitated by the peculiar potassic nature of the magmas feeding the peri-Tyrrhenian potassic Quaternary volcanoes (e.g., Peccerillo, 2017), which generated products bearing K-rich minerals (e.g., sanidine and leucite), ideal for direct $^{40}Ar/^{39}Ar$ dating. Furthermore, with regard to the explosive activity, the great deal of tephrostratigraphic research on distal stratigraphic successions, i.e., located tens to thousands of km from the volcanic sources, substantially contributed to increase the knowledge on the recurrence, temporal extent and dynamics of explosive activity, providing integrative information that are difficult to detect in the proximal areas (e.g., Monaco et al., 2021, 2022a; Leicher et al., 2023). Finally, the last twenty years have seen an increase in research aimed at evaluating volcanic risk in areas of active volcanism, with particular reference to hazard assessment based on knowledge of past eruptive history (e.g., Cioni et al., 2003; Lirer et al., 2010; Del Negro et al., 2013; Bevilacqua et al., 2022 and references therein).

Given these premises, it seemed important to gather the contributions of various members of the Italian volcanological community in order to present an updated review on the Italian Quaternary volcanoes. To emphasize the volcanological aspects over the petrographic-magmatological ones as companion papers we deliberately decided to not adopt the division into magmatic provinces (see Conticelli et al., 2010, 2015a; Peccerillo, 2017; and references in these papers and book) but to list the Italian Quaternary volcanoes in a simple and neutral geographical way, starting the from Tuscany and then moving southward on to those located in Lazio, Campania, Tyrrhenian Sea and Sicily (Fig. 1).

For each volcanic area or volcano dealt with, a summary of the eruptive history is proposed, focusing on the dynamics in relation to time, with emphasis on the most recent researches, and shortly mentioning the chemistry of the products. For the active volcanic areas, particularly those exposed to higher risk, information is also given on the state of hazard and risk assessment research.

We hope that this paper will be a valid support for the studies of the Quaternary community, which very often has to deal with the eruptive history of the Italian volcanoes and for which it is extremly necessary to have an agile, exhaustive and up-to-date general framework.

2. GEODYNAMIC FRAMEWORK OF QUATERNARY VOLCANIC ACTIVITY IN ITALY (S.C., B.G., P.P.)

The Tyrrhenian Sea is one of the most geologically complex areas in the world (Faccenna et al., 2001b; Cavazza et al., 2004; Beltrando et al., 2010). The reasons for this complexity are mainly linked to its role in the convergence of the African and European plates, which starting from the Cretaceous to the Present time, has led to the formation of two orogenic belts (Alpine and Apennine Chain) and several back-arc basins of progressively younger age moving from W to E in the western Mediterranean (Sartori, 2003). Magmatic processes are integral part of this geodynamically complex region and punctuated all stages of its evolution. In the initial stages this activity was concentrated in the Alps (e.g., Conticelli et al., 2009b, 2010; Alagna et al., 2010) along the margins of the colliding plates and, to a lesser extent, within the Africa-Adria plate. The last phase of this complex geodynamic magmatic pattern began in the Upper Eocene-Oligocene and is still on-going. This is coeval with the formation of the Apennine-Maghrebian chain, and the opening of the western Mediterranean basin (Lustrino et al., 2011; Di Capua et al., 2016), and has therefore been referred to as "the Apennine magmatism" (Peccerillo, 2017). Opening of slab tears and incipient slab detachment also occurred in the late evolution of the Apennine orogen, favouring the channelling of intra-plate volcanism that mixed with the orogenic one during the final stage of Vulture volcanism (e.g., Faccenna et al., 2001a, 2001b, 2004; Avanzinelli et al., 2008, 2009; Conticelli et al., 2009b). The Apennine orogen has been subject to back-arc extension since the Late Miocene due to the rapid slab roll-back of the Calabrian block (Malinverno & Ryan, 1986), which has produced two major extensional basins in the abyssal plain of the Tyrrhenian Sea, the Vavilov and the Marsili basins, located at north-west and south-east portions of the Southern Tyrrhenian basin, respectively (Marani and Gamberi, 2004). The opening of the Marsili back-arc basin also coincides with the formation of the Aeolian volcanic arc, whose spatial evolution is linked to the decoupling and roll-back of the slab beneath the Calabrian Arc (Loreto et al., 2020). At the back of the eastward migrating Apennine orogen intense extension occurred since the Upper Miocene (Jolivet et al., 1998, and references therein). In the central Tyrrhenian margin, the onset of extensional tectonics can be dated at Late Miocene, the age of syn-rift sedimentary sequences filling in the extensional basins located between the Tolfa-Ceriti and Roccamonfina volcanoes (Mattei et al., 2010). The Tyrrhenian margin of southern Italy, on the counterpart, has been mainly characterized by Plio-Quaternary back-arc extension, with the development of NW-SE normal faults (Acocella and Funiciello, 2006); coeval NE-SW transverse systems are also present. Despite the overall NW-SE alignment of the Quaternary volcanoes along the Tyrrhenian margin, most of them formed where NW-SE normal faults intersect transverse NE-SW tectonic lineaments which, thanks to the higher vertical depth with respect to the NW trending features, represent a preferential structure for magma uprising and fluids upwelling (Acocella and Funiciello, 2002). The volcanic edifices consist of poli-phase calderas and large ignimbrite plateaus, which identify vast areas with negligible topographic relief (Vulsini, Sabatini, the Vulcano Laziale - the early edifice of the Colli Albani volcano, Campi Flegrei) and stratovolcanoes with summit calderas (e.g., Vico, Faete - the intracaldera edifice of Colli Albani, Roccamonfina and Vesuvius).

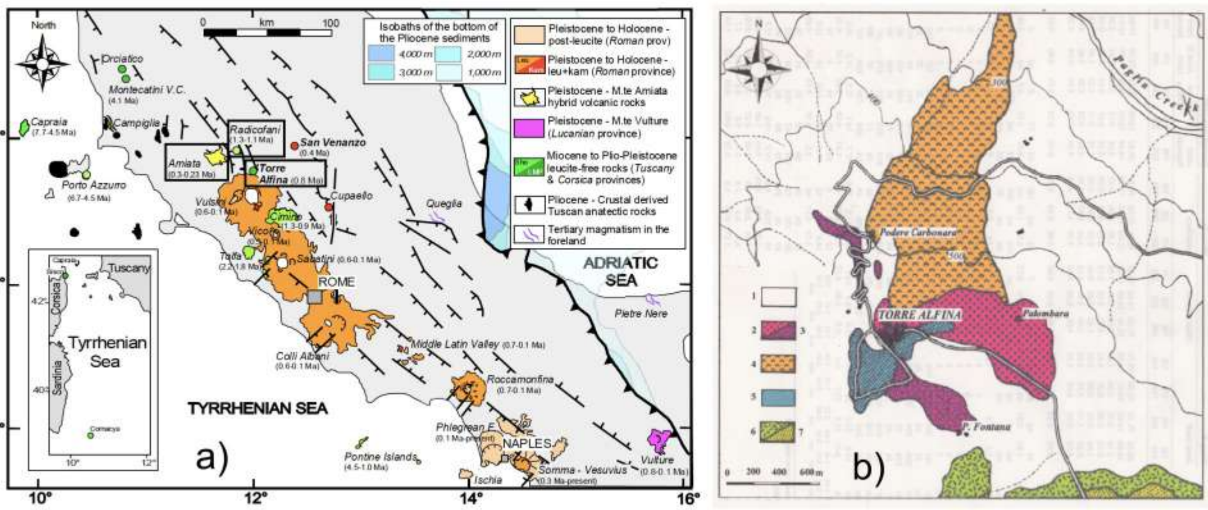


Fig. 2 - a) Distribution of ultrapotassic rocks in the Central Italian region revealing the location the Torre Alfina, Radicofani and Monte Amiata volcanoes (spots within the large rectangular shapes) and the distribution of the major magmatic potassic and ultrapotassic events. Red spots and areas (kamafugites) and dark green spots (lamproites) mark the location sites of the most extreme endmembers of the orogenic magmatic activity. Modified after Conticelli et al. (2010, 2013, 2015a) and Günther et al. (2023); b): Geological sketch map of the Torre Alfina volcano. Legend: 1) Cretaceous carbonaceous and argillaceous sedimentary rocks; 2) blocky olivine-latite lava flows, type B; 3) low viscosity olivine-latite lava flows, type B; 4) low-viscosity olivine-latite lava flows, type A; 5) lapilli and scoriae; 6) leucite-tephrite to leucite-phonolite Vulsini lavas; 7) leucite-phonolite to leucite-trachyte Vulsini pyroclastic rocks. Redrawn after Conticelli (1998) and Avanzinelli et al. (2017).

3. THE CENTRAL ITALY QUATERNARY VOLCANOES

3.1. Tuscany (S.C., J.P.S.-B.)

3.1.1. General background

The oldest Potassic to Ultrapotassic volcanic rocks of the Central Mediterranean region are found on the western margin of the Tyrrhenian basin, distributed N-S (Fig. 2), in Corsica at Sisco (14.1 Ma; Civetta et al., 1978), at Capraia Island (7.5-4.6 Ma; Gasparon et al., 2009) and at Cornacya Seamount (12.6 Ma; Mascle et al., 2001). These are leucite-free, lamproite-like, ultrapotassic rocks associated with shoshonitic and high-K calc-alkaline volcanic rocks (Conticelli et al., 2009a). Following the rollback of the slab and the consequent opening of the Tyrrhenian basin, the volcanism shifted diachronically eastward, from Miocene to Pliocene, from Corsica to Tuscany (Fig. 2) (e.g., Avanzinelli et al., 2009; Conticelli et al., 2010, 2015a).

In Tuscany, lamproite like, leucite-free ultrapotassic rocks and associate shoshonitic and calc-alkalic volcanic rocks of ultimate mantle origin are distributed on both sides of the western shoreline of the Italian peninsula (Fig. 2a; e.g., Peccerillo et al., 1987; Conticelli & Peccerillo, 1992; Conticelli et al., 2009a, 2010, 2015; Peccerillo, 2005, 2017), and in some cases these volcanic rocks are intimately associated, in space, but not in time, with crustal-derived rhyolitic to granitic rocks (Peccerillo et al., 1987; Conticelli et al., 2002, 2004, 2010; Conte & Dolfi, 2002).

The oldest ultrapotassic to high-K calc-alkalic igneous rocks of the so called "*Tuscan Magmatic Province*" (Conticelli & Peccerillo, 1992) are found at Elba Island, in the archipelago between Italian Peninsula and Corsica with a $^{40}Ar/^{39}Ar$ age of 5.89 Ma (Conticelli et al., 2001). These were followed by the emplacement, in the mainland, of the Val d'Era lamproites made of the hypabissal minette of Montecatini Val di Cecina, emplaced at 4.1 Ma (K/Ar; Borsi et al., 1967), and lamproite-like "orendite" of Orciatico (Conticelli et al., 1992), of the latitic dykes at Campiglia (Conticelli et al., 2004), and in the Latium region of the volcanic products of the Tolfa-Manziana-Ceriti dome complexes (Fig. 2a) between 4.3 and 1.9 Ma (e.g., Evernden & Curtis, 1965; Fazzini et al., 1972; Bigazzi et al., 1979; Lombardi et al., 1974; Villa et al., 1989).

The Quaternary volcanic rocks are instead found at Radicofani, Torre Alfina, Monte Cimino, Ponza Island (Monte La Guardia) and in drills in the Campania region. These leucite-free ultrapotassic, lamproite-like, to shoshonitic and calc-alkalic rocks were emplaced between 1.3 and 0.85 Ma (e.g., Evernden & Curtis, 1965; Nicoletti, 1969; Pasquarè et al., 1983; Sollevanti, 1983; D'Orazio et al., 1991; Laurenzi et al., 2015; Conte et al., 2016; Nappi et al., 2022).

3.1.2. Torre Alfina monogenetic volcano

The Torre Alfina Volcano (Fig. 2b) is located a few kilometers north of the Vulsinian district (Fig. 2; 42° 45'19"N-11°56'41"E) and consists of a few tiny lava flows erupted from a small volcanic centre, and a small neck cropping out North of Torre Alfina castle. Torre Alfina is a monogenetic volcano characterised mainly by lava flows and minor Strombolian activity (Conticelli, 1998). The Torre Alfina volcanic rocks rest directly on the sedimentary substratum of the Monte Cetona horst.

The Torre Alfina lava flows were erupted at the top of the hill and flowed partly into the canyon formed by the Paglia river. Field characteristics indicate that they had low viscosities, and some lava tunnels may be recognised in the eastern part of the vent. Two slightly different types of lava can be also recognised. They were emplaced contemporaneously. They have different colours, vesiculation, and porphyritic indexes. The most mafic lava (type-A, Fig. 2b) is dark grey, almost aphyric, and contains abundant green ultramafic microxenoliths 2-4 cm in diameter (Conticelli & Peccerillo, 1990), as well as a few lower crustal xenoliths of variable size (5-10 cm) and nature (Orlando et al., 1994). Type-B lava is the most widely represented at Torre Alfina volcano. It is dark grey to light grey in color, with a variable porphyritic index and vesiculation, showing some flattened vesicles of a few centimeters in length. Type-B lavas contain abundant crustal xenoliths of variable size (2-30 cm), some large (3-20 cm) magmatic mica-rich inclusions, and rare green ultramafic xenoliths. Torre Alfina lavas are leucite free ultrapotassic rocks of lamproitic nature of ultimate mantle origin generated by low degrees of partial melting of a veined mantle wedge at low X_{CO2} conditions (Conticelli et al., 2015a).

3.1.3. Radicofani monogenetic volcano

The Radicofani volcano is a monogenetic volcano lying within the central portion of the Radicofani basin, few kilometers North of the Torre Alfina volcano (Fig. 2b). The Radicofani basin represents the southernmost branch of the Siena-Radicofani graben, one of the most important post -orogenic extensional tectonic features of the Northern Apennine (e.g., Pasquaré et al., 1983; Acocella et al., 2002; Bonini & Sani, 2002). The Radicofani monogenetic volcano is formed by a 90 m-high well-preserved volcanic neck, with remnants of a cinder cone and of a lava lake at its top, and by several lava flows (Fig. 3).

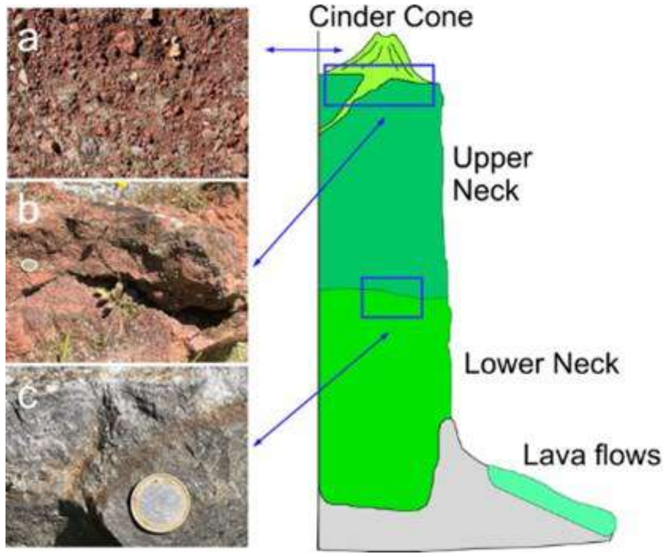


Fig. 3 - (a) Scoria and lapilli of the cinder cone containing the lava lake at the roof of the neck; (b) Reddish shoshonitic lava forming the lava lake at the roof of the neck; (c) Oxidised septum that separates shoshonitic lava of the upper neck from the high potassium basaltic andesitic lava of the lower neck (redrawn after Conticelli et al., 2011; Avanzinelli et al., 2022).

The lava flows are scattered around the neck and lie on top of the Pliocene marine sediments (Liotta, 1996; Disperati & Liotta, 1998; Ghinassi & Lazzarotto, 2005; Conticelli et al., 2011). The original volcanic edifice was made up by a cinder cone a few hundred meters high. Today only a thin layer of red scoriae is preserved on the edge of the top of the neck, just West of the homonym castle. The top of the neck is made up of red vesicular lava that becomes dense and grey downward (Fig. 3). Columnar jointing is present in the middle and lower portions of the neck in grey and dark grey aphyric to sub-aphyric lavas. Oxidised septa dividing the different portions of the neck can also be observed. Rocks of different portions of the volcano yield radiometric ages in the range between 1.315 Ma and 1.08 Ma (e.g., Barberi et al., 1971; Pasquaré et al., 1983; D'Orazio et al., 1991), but the uncertainty in the age determination is relevant.

The Radicofani volcanic rocks have significant compositional variations from basaltic andesite (high-K calc-alkaline), at the bottom of the neck to ultrapotassic shoshonite for the lava-lake lavas, scoria and lava flows. The overall volcanic rocks show aphyric to sub-aphyric textures with small euhedral phenocrysts of olivine set in a microcrystalline matrix made of clinopyroxene, sanidine, ilmenite, and magnetite. Rare microliths of orthopyroxene are found in the groundmass of calc-alkaline samples. Chemical and isotopic variations along the necks are thought to be a primary feature due to incremental partial melting of a veined mantle wedge under conditions of low X_{CO2}, favouring the genesis of silica-saturated, leucite-free, ultrapotassic magmas (Conticelli et al., 2011).

3.1.4. The Monte Amiata: a Quaternary “hybrid” volcano

The Monte Amiata volcano is a small linear Late Pleistocene volcano located some tens of kilometers North of the Vulsini district (Fig. 1) and few kilometers west of Radicofani monogenetic volcano (Figs. 2a and 4). It is found where the Roman and Tuscany magmatic provinces overlap in space but not in time (Conticelli et al., 2004, 2010).

The Monte Amiata volcano produced a sequence of several viscous lava flows and exogenous domes outpoured from NNE-SSW tectonic structures (Fig. 2a), both hosting abundant cognate magmatic enclaves (e.g., Ferrari et al., 1996; Marroni et al., 2015). The volume of magma outpoured is smaller than the amount of magma erupted in Roman volcanoes, but consistently larger than that produced by Tuscan volcanoes (Conticelli et al., 2009a; 2010). According to Laurenzi et al. (2015) the onset of magmatic activity was at 305 ka with final lavas emplaced at 230 ka. Lavas and enclaves are characterised by high-K calc-alkaline to shoshonitic affinities ranging in composition with time from trachydacites to trachyte, latite, and olivine latite, with a progressive increase of the amounts of mafic enclaves. Monte Amiata rocks, lavas and enclaves, are leucite free, with phenochrysts of olivine, confined to the most mafic terms, biotite, plagioclase, sanidine, clinopyroxene and orthopyroxene, set in a microcrystalline to glassy matrix. Fine-grained magmatic enclaves range from porphyritic to aphyric and invariably display chilled margin texture, ranging in compositions from trachybasaltic to shoshonitic and latitic. Petrographic and compositional characteristics of fine-grained magmatic enclaves suggest that they were molten at the time of inclusion by the host trachytic magma. Their chemical, mineralogical, and isotopic characteristics are strongly suggestive of genesis by mixing/mingling between a resident trachydacitic mag-

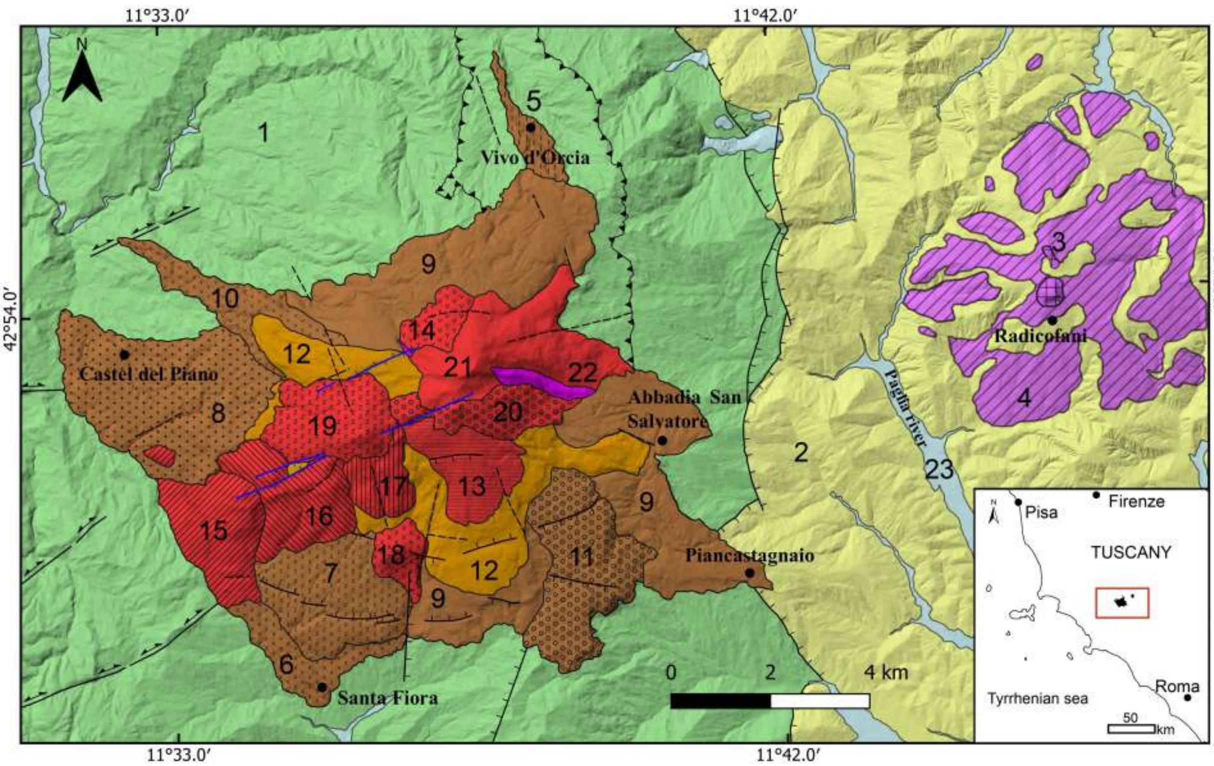


Fig. 4 - Geological sketch map of the Amiata region. Legend: 1) sedimentary sequences of the Ligurid units; 2) Pliocene marine sediments; 3-4) Radicofani volcano, neck and dismembered lavas, respectively; 5-14) Monte Amiata Volcano - 5-6) Basal trachydacitic lava flows (6); 7-11) trachytic domes and lava flows; 13) latitic lava flows; 14) olivine latitic lava flows; 15) Quaternary alluvial deposits (redrawn after Conticelli et al., 2015a,b; Laurenzi et al., 2015; Marroni et al., 2015).

ma, derived by crystal fractionation from a high-K calc-alkaline basaltic-andesitic magma (Tuscan type), similar in composition to the Radicofani magma and a younger Roman-type, silica-undersaturated, basanitic magma (Conticelli et al., 2015b, and references therein).

3.1.5. The other Leucite-free volcanoes in Latium and Campania

Other leucite-free ultrapotassic associated to shoshonitic and calc-alkalic volcanic rocks of ultimate mantle origin were emplaced along the Tyrrhenian border of the Italian peninsula in a period of time between 2 and 1 Ma (Fig. 2) at Monte Cimino volcanic complex, in Northern Latium underlying the Vico Volcano (Conticelli et al., 2013), at Tolfa-Maziana-Cerite volcanic complexes, in Central Latium underlying the Sabatini volcanoes (Bertagnini et al., 1995), offshore in the Pontine archipelago, in Southern Latium (Conte & Dolfi, 2002; Cadoux et al., 2005), and in the Neapolitan area, in the Central Campania, beneath the Campi Flegrei volcanoes (Albini et al., 1980; Barbieri et al., 1979; Brocchini, 1999). In most of these localities the older leucite-free igneous rocks preceded by a few hundred thousand years the younger strongly silica-undersaturated, melilite- to leucite -bearing Roman volcanic rocks (Conticelli et al., 2010, 2015a).

3.2. Latium and northern Campania

(D.M.P., S.C., J.P.S.-B., M.G. B.G., G.G., F.M., G.S.)

3.2.1. General background

The Quaternary (Middle Pleistocene to present) ultrapotassic magmatism of Central Italy fits into the geodynamical context of the Tyrrhenian back-arc extension related to the NE retreat of the west-directed Adriatic slab (Malinverno & Ryan, 1986; Conticelli & Peccerillo, 1992; Faccenna et al., 2001a; Acocella & Funiciello, 2006). This originated a NW-SE trending volcanic belt extending between the Tyrrhenian Sea coast and the Apennine orogen from southern Tuscany, through Latium and Campania. Defined as the Roman Comagmatic Region since Washington (1906), this volcanic region is known worldwide for its peculiar potassic and strongly silica-undersaturated and melilite- to leucite-bearing ultrapotassic magma: feeder magmas display a wide compositional range and multiple differentiation trends, ranging from K-basalts to K-foidites, phonolites and trachytes, and even K-rhyolites. Based on isotope geochemistry, the Roman, Ernici-Roccamonfina and Campanian districts are distinguished (Peccerillo, 2005, 2017; Conticelli et al., 2010).

Here, we focus on the Roman Province volcanoes, i.e., from NW to SE, Vulsini, Vico, Sabatini and Colli Albani, as well as on the Volsci Volcanic Field (formerly erroneously defined Ernici volcanoes) and Roccamonfina volcano, which belong to the Ernici-Roccamonfina districts (Fig. 5). These volcanoes show a full spectrum of

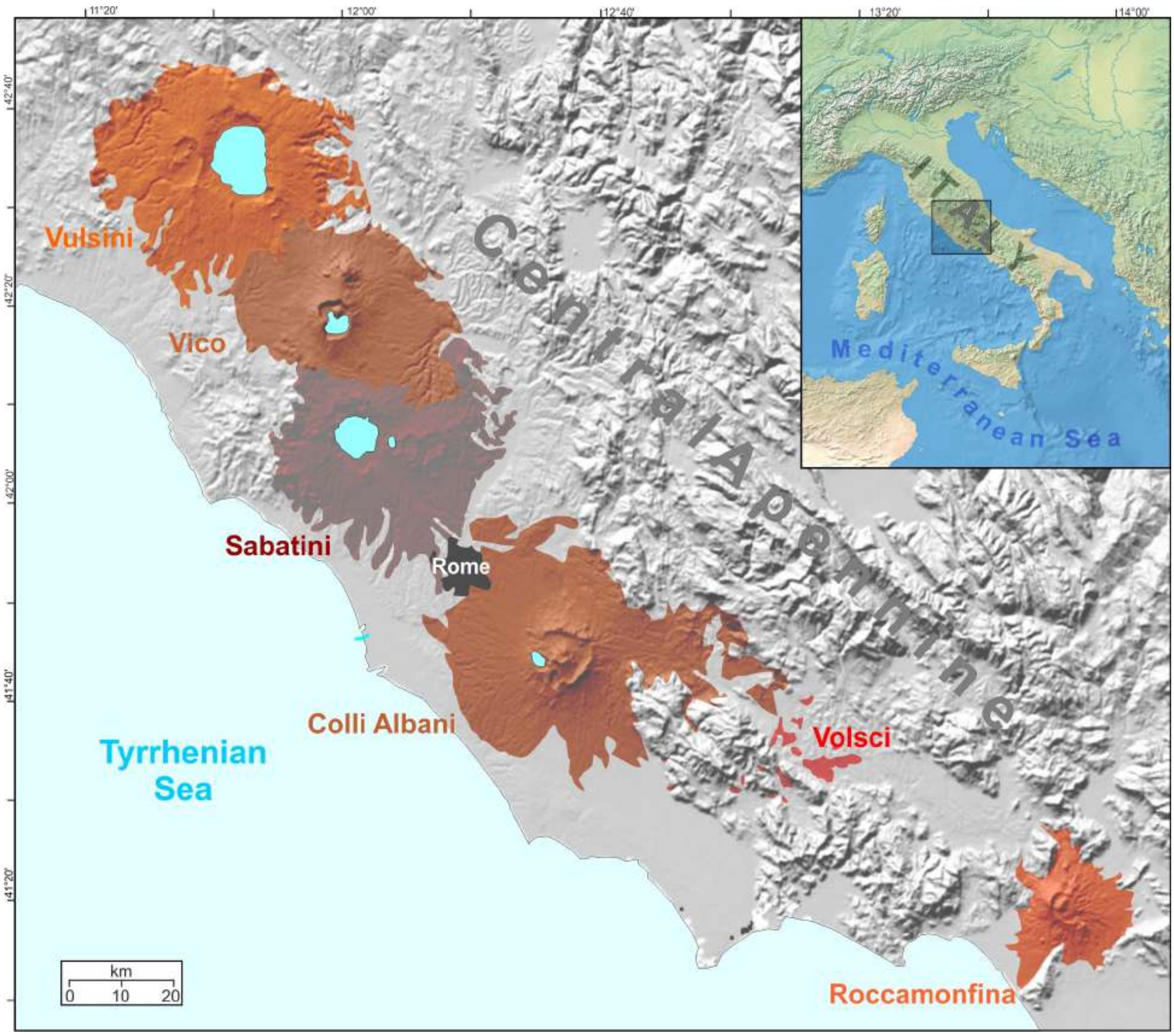


Fig. 5 - Location and geological sketch of the Quaternary peri-Tyrrhenian potassic volcanoes of central Italy.

eruptive dynamics and related morphologies, ranging from small-scale Strombolian and phreatomagmatic events from monogenetic centers (scoria cones, maar-tuff rings and tuff cones), with associated effusive activity, to highly explosive Plinian and pyroclastic flow-forming events, often related to caldera collapses. Major volcanic districts with a dominant areal character (i.e., Vulsini and Sabatini), characterized by multiple caldera depressions and networks of monogenetic eruptive centers, alternate with essentially central volcanoes, such as stratovolcanoes with summit calderas and subordinate eccentric activity (i.e., Vico and Roccamonfina), or polygenetic caldera volcanoes with ignimbrite plateaux (i.e., Vulcano Laziale in the Colli Albani district). In some cases, central volcanic edifices with summit calderas, such as Latera and Montefiascone (Vulsini), Sacrofano (Sabatini) and Faete (Colli Albani), superimposed in time and space with areal volcanism. Quite peculiarly, the Volsci Volcanic Field consists of small-scale, isolated or locally clustered, eruptive centers.

The onset of potassic volcanism is documented by the oldest exposed products in the Volsci Volcanic Field that date back to at least ca. 760 ka (Boari et al., 2009b; Cardello et al., 2020; Marra et al., 2021) (Fig. 6). However, evidence of even earlier activity in the Roman area is provided by ca. 750-810 ka tephra layers found in the Tiber River delta (Marra et al., 2014; Fig. 6). Since ca. 600 ka, the Vulsini, Sabatini, Colli Albani, Volsci and Roccamonfina volcanoes were all simultaneously active (Fig. 6) (e.g., Marra et al., 2003; Sottili et al., 2004; Rouchon et al., 2008; Boari et al., 2009b; Conticelli et al., 2010; Palladino et al., 2010; Soligo & Tuccimei, 2010), followed at ca. 415 ka by the onset of volcanic activity at Vico (Pereira et al., 2020).

3.2.2. Vulsini Volcanic District (ca. 590-111 ka)

The northernmost and largest (ca. 2200 km^2) volcanic district in the Roman Province developed along the southern termination of the NNW-SSE trending Miocene-Pliocene Siena-Radicofani Graben from the superposi-

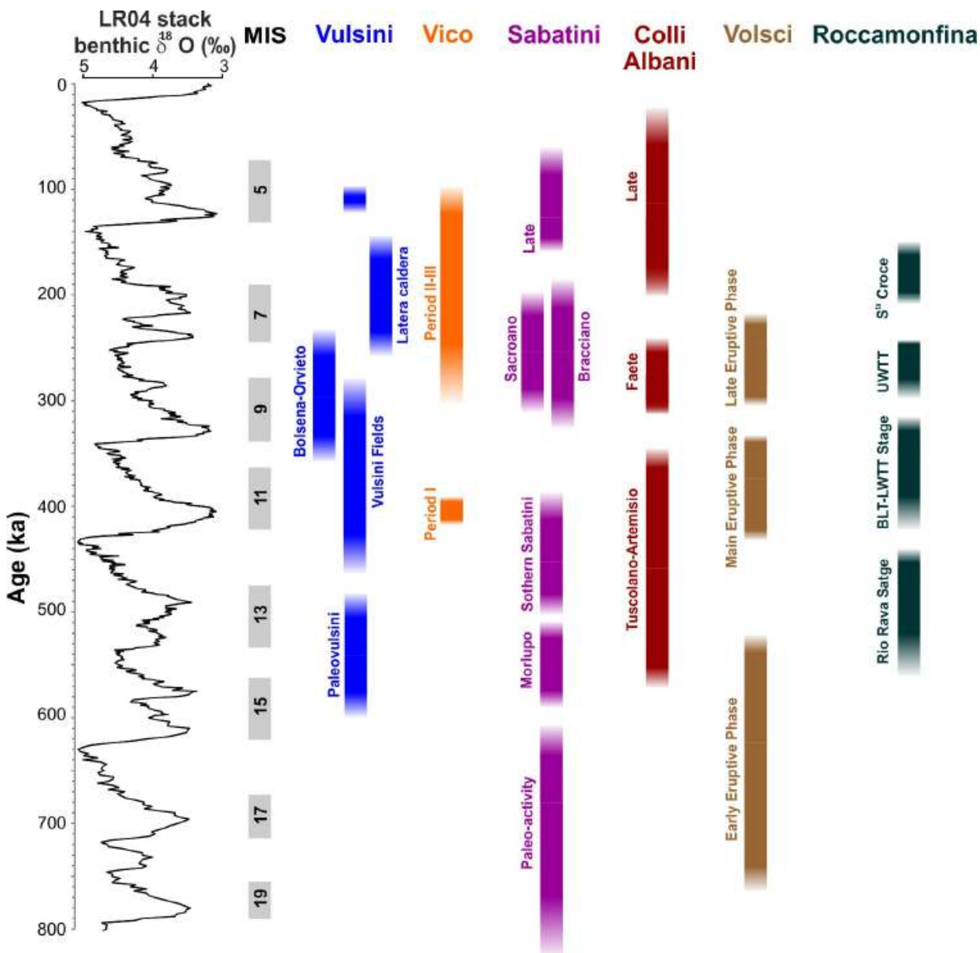


Fig. 6 - Temporal distribution of the Middle Pleistocene volcanic activity from the peri-Tyrrhenian potassic volcanic systems, plotted against the LR04 benthic stack record (Lisiecki & Raymo, 2005). Data source: Vulsini: Palladino et al. (2010), Marra et al. (2020a); Vico: Perini et al. (2004), Pereira et al. (2020); Sabatini: Sottili et al. (2010), Marra et al. (2014, 2020b); Colli Albani: Marra et al. (2009); Volsci: Boari et al. (2009), Centamore et al. (2010), Marra et al. (2021); Roccamonfina: Giannetti (1996a, 1996b), Giannetti & De Casa (2000), Rouchon et al. (2008), Scaillet et al. (2008).

tion of five major volcanic complexes (or lithosomes), partially overlapping in space and time: Paleovulsini, Vulsini Fields (formerly defined as Southern Vulsini), Bolsena-Orvieto, Montefiascone and Latera (Vezzoli et al., 1987; Palladino et al., 2010 and reference therein; Figs. 7 and 8).

The most prominent volcano-tectonic feature in the district is the wide caldera depression hosting Lake Bolsena, which is intersected, respectively to the W and SE, by the Latera and Montefiascone central volcanoes, featuring collapse calderas (Nappi et al., 1991; Acocella et al., 2012). The whole area is dotted with several tens of monogenic cones and craters, either isolated (e.g., Lagaccione maar, tuff cones of Bisentina and Martana islands in Lake Bolsena) or clustered (e.g., the Valentano scoria cones on the SE rim of Latera caldera) or located along volcano-tectonic lineaments (e.g., the N-S scoria cone alignment along the eastern rim of Bolsena caldera).

Overall, volcanic activity spanned a broad spectrum of eruptive styles, intensities and magnitudes, ranging from small-scale explosive (Strombolian and hydromagmatic) and effusive eruptions from monogenetic centers, up to Plinian and pyroclastic flow, caldera-forming, events (Sparks, 1975; Conticelli et al., 1987; Nappi et al., 1994; Palladino & Valentine, 1995; Palladino & Agosta, 1997; Palladino & Simei, 2005; Palladino et al., 2014). Figure 9 shows an example of interbedded volcanic units from different source vents and their unconformity boundaries. In the following, the five lithosomes are briefly described:

1) the Paleovulsini lithosome groups the early eruptive products of Vulsini (ca. 590 to <490 ka; Nicoletti et al., 1981; Vezzoli et al., 1987; Cioni et al., 1989; Nappi et al., 1995), which consist of trachytic welded pyroclastic-flow deposits (basal ignimbrites or Nenfri, ca. 550-505 ka) and associated Plinian pumice fall horizons exposed at the periphery of the district on top of the sedimentary substrate. An early collapse caldera possibly related to the basal ignimbrites, inferred from the Bolsena lake bathymetry and deep drillings, controlled the location and style of the following eruptive activity, which followed as areal volcanism (Vulsini Fields lithosome);
2) the Vulsini Fields (former Southern Vulsini complex;

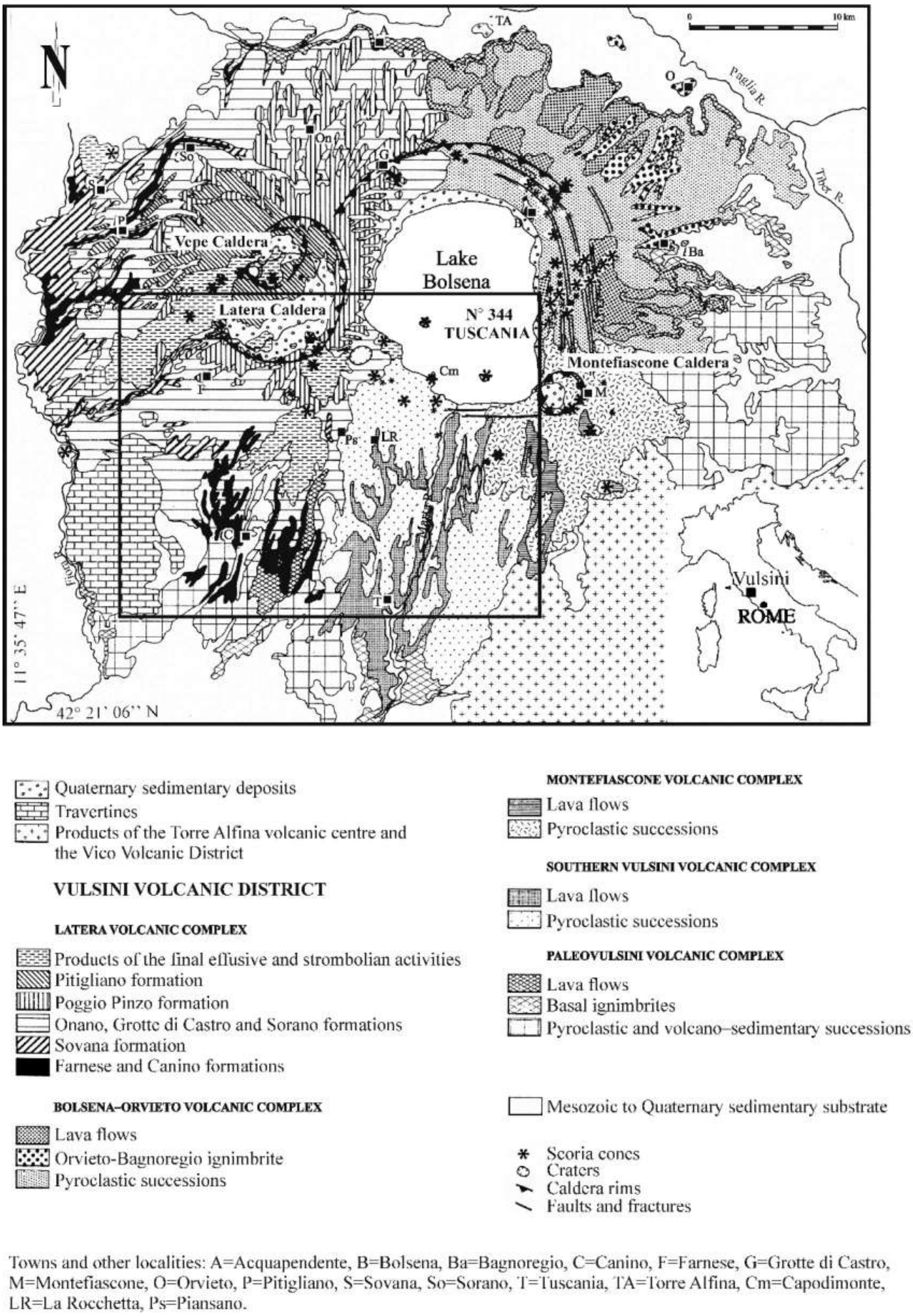


Fig. 7 - Geological sketch map of the Vulsini Volcanic District (modified after Palladino et al., 2010). The boundaries of the Sheet No. 344-Tuscania of the 1:50,000 Geological Map of Italy (CARG project) are also shown. Towns and other localities: A, Acquapendente; B, Bolsena; C, Canino; F, Farnese; G, Grotte di Castro; M, Montefiascone; O, Orvieto; P, Pitigliano; S, Sovana; So, Sorano; T, Tuscania; TA, Torre Alfina; Cm, Capodimonte; LR, La Rocchetta; Ps, Piansano.

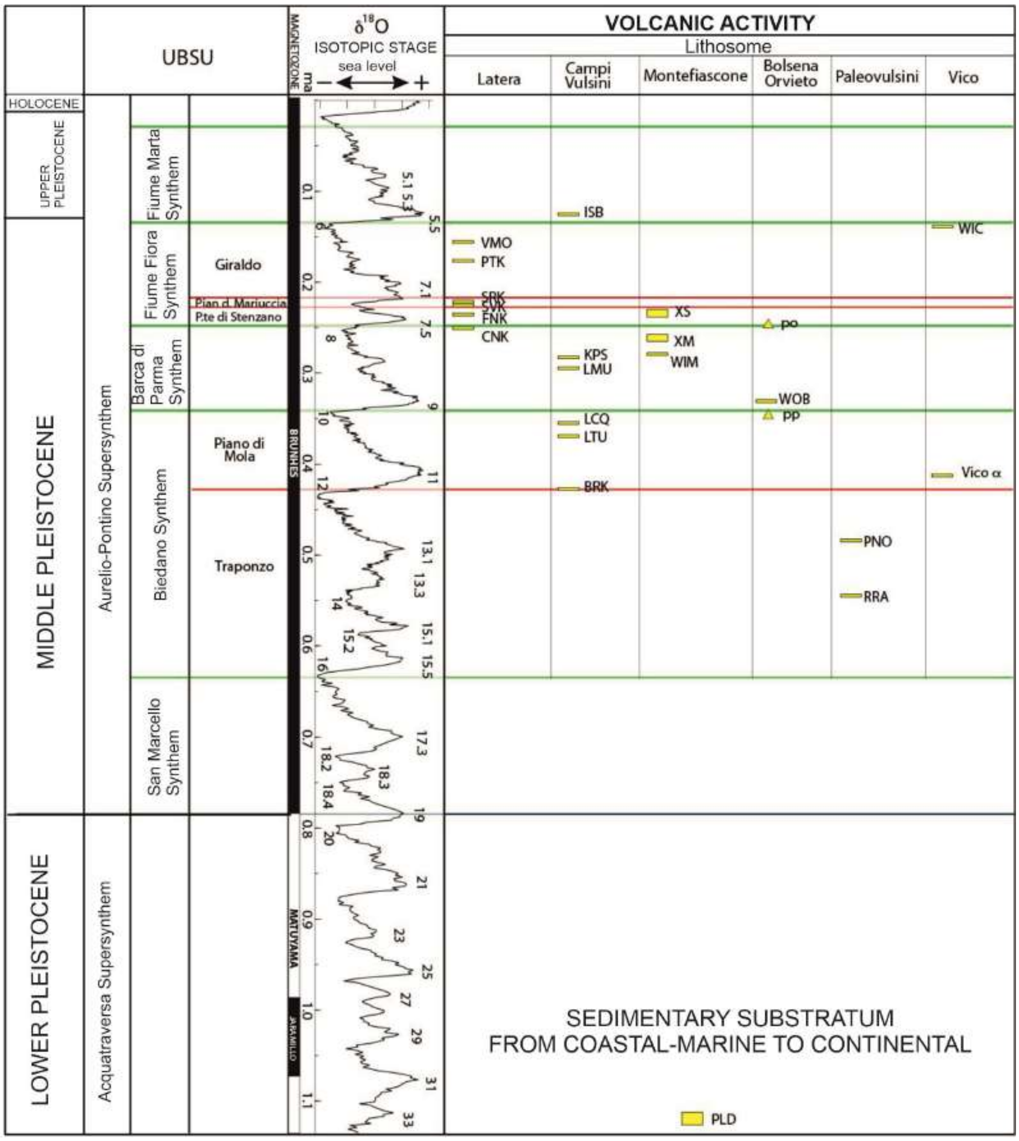


Fig. 8 - Sketch of the legend of sheet 344-Tuscania (after Palladino et al., 2010, modified), where representative lithostratigraphic volcanic units are grouped into lithosome units and organized in the framework of the unconformity-bounded stratigraphic (UBS) units recently defined for the volcanic and Tyrrhenian coastal areas of northern Latium. Note that major UBS unit boundaries correlate to even-numbered oxygen isotope stages, corresponding to low-stands of sea level ($\delta^{18}O$ curve after Shackleton et al., 1990; Shackleton, 1995).

Vezzoli et al., 1987) is a composite lithosome consisting of a gently sloping volcanic plateau, extending throughout the whole district, which developed by long-lasting (ca. 490-111 ka) effusive and subordinate explosive activity from a network of scattered eruptive centers. Volcano-tectonic lineaments related to the early Paleovulsini caldera might have favoured the eruption of poorly differentiated, small magma batches from multiple source vents around the present Lake Bolsena depression. Although this areal volcanism largely preceded the onset of central explosive activities at Latera and Montefiascone at ca. 280 ka (Nappi et al., 1995; Brocchini et al., 2000), it was also contemporaneous and even followed the Montefiascone and Latera explosive climax until the waning phases of activity in the district. In particular, the Monte Bisenzio spatter cone represents the most recent volcanic product dated so far at Vulsini (ca. 111 ka; Marra et al., 2020a). In lack of evidence of major caldera-forming eruptions that could explain the size of the Bolsena depression, Walker (1984) and Cole et al. (2005) classified it as a downsag caldera end-member. The relationships among eruptive sources and volcano-tectonic lineaments may suggest that the prolonged magma drainage during the Vulsini Fields activity was accompanied by incremental caldera growth that extended the early Paleovulsini caldera collapse and eventually resulted into the wide depression hosting Lake Bolsena (Acocella et al., 2012);

3) the Bolsena-Orvieto lithosome includes lava and pyroclastic successions that are mainly exposed in the northeastern sector of Vulsini, from Lake Bolsena to the Tiber Valley, including a series of Plinian fall deposits

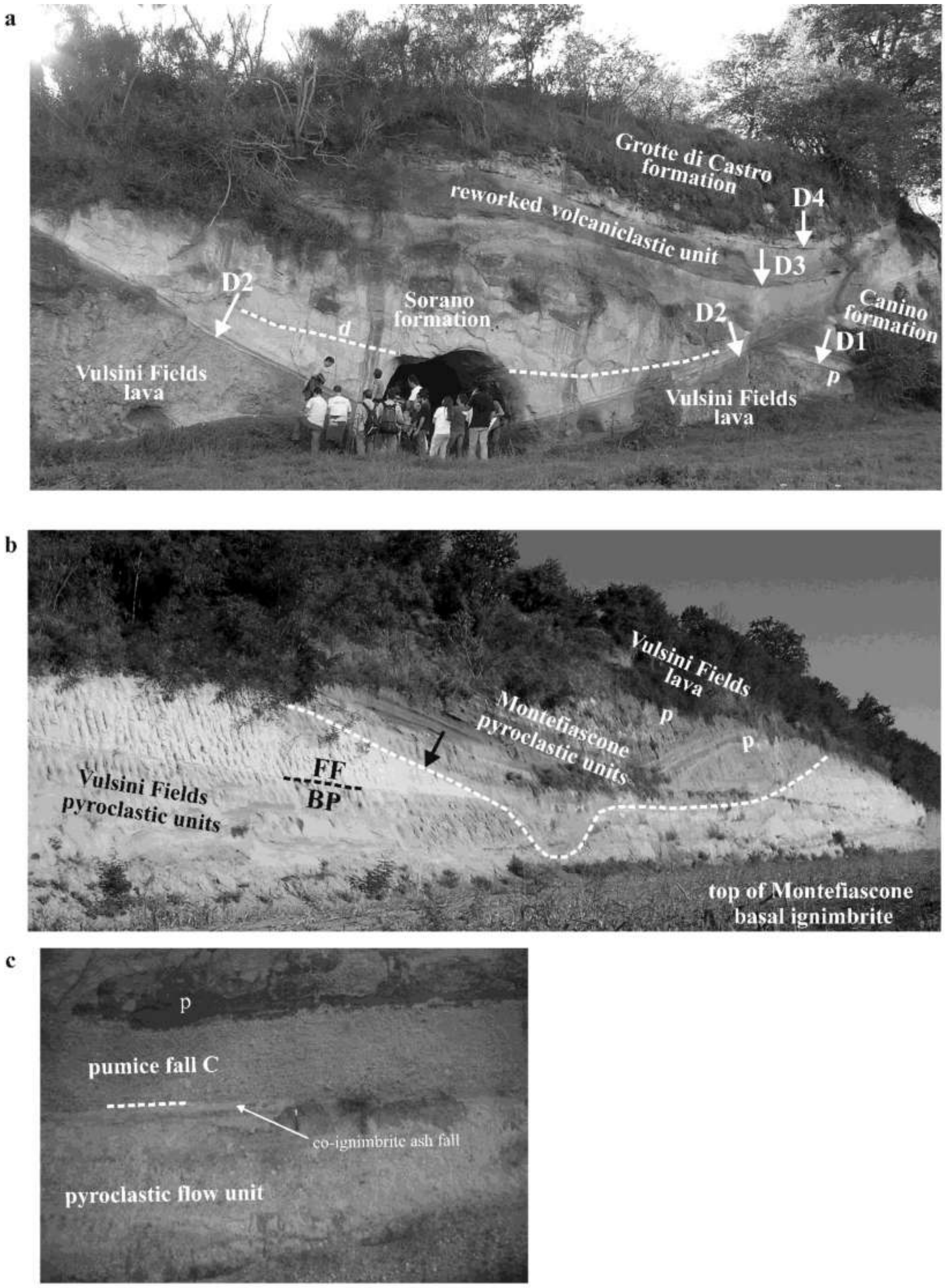


Fig. 9. Field pictures of the Vulsini volcanic successions. (a) La Rocchetta locality (6 km SW of Capodimonte): major pyroclastic units of Latera (i.e., Canino, Sorano, Grotte di Castro formations) on top of a lava flow from the Vulsini Fields and intervening stratigraphic discontinuities, such as paleosols (e.g., D1, p) and high-relief erosional surfaces (e.g., D2, D3, D4), related to UBS boundaries of different rank (see Fig. 8). Note the U-shaped channel (D2) mantled by a thin ash-fall layer and filled by pyroclastic current deposits of the Sorano formation (flow direction toward the page; d-example of flow unit boundary). (b) 5 km SW of Montefiascone: interbedded Vulsini Fields and Montefiascone volcanic successions, showing the slightly undulating erosional surface bounding the Barca di Parma (BP) and Fiume Fiora (FF) synthems, and local stratigraphic discontinuities representative of temporal hiatuses in the eruptive activity, as evidenced by steep erosional channels (black arrow) or paleosols (p). (c) Close-up of the upper part of the Canino formation at La Rocchetta, showing the depositional boundary between temporally closely spaced pyroclastic flow and fall units, topped by the Canino-Farnese inter-eruptive paleosol.

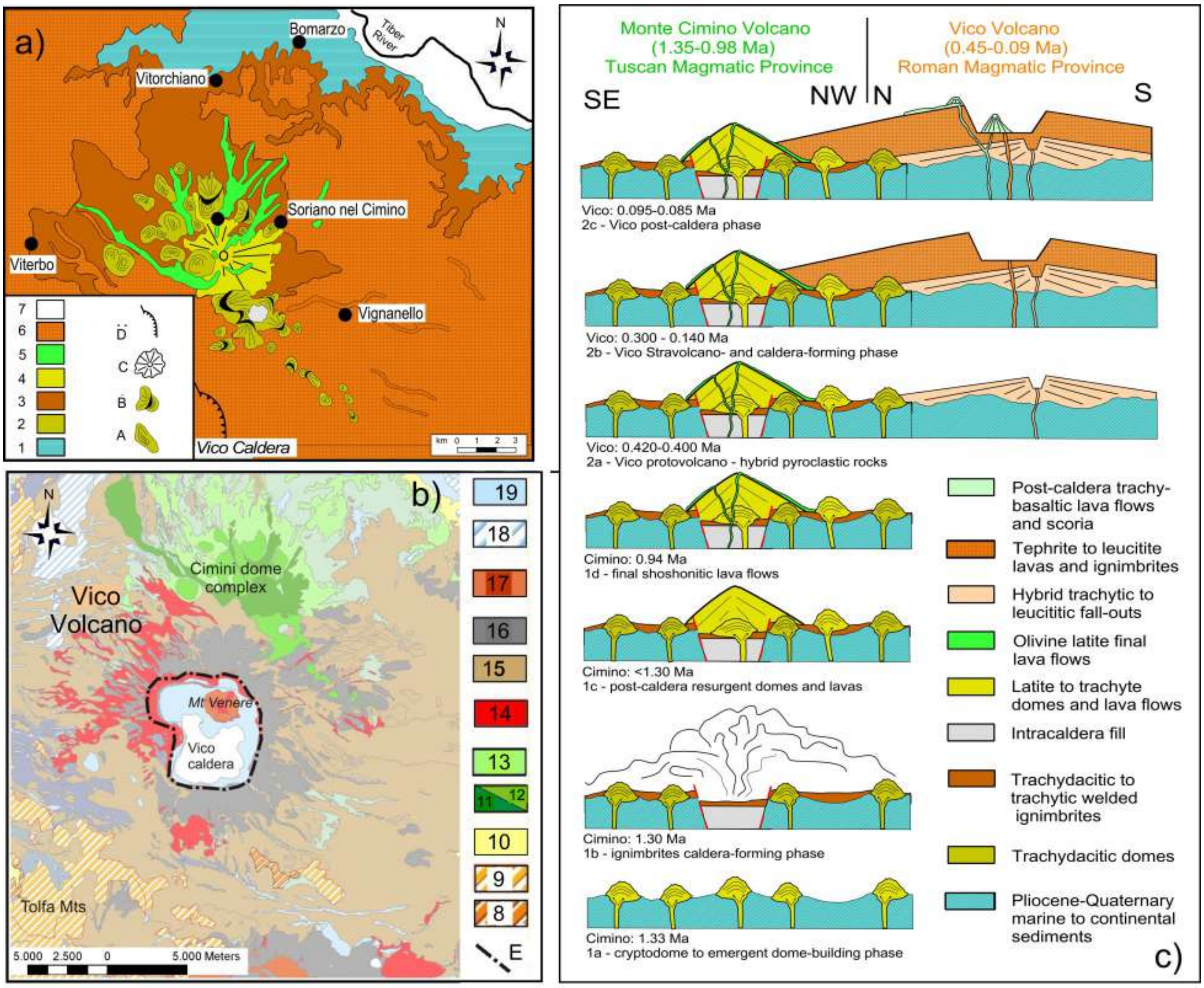


Fig. 10 - (a) Geological sketch map of the Cimino Volcanic Complex (Tuscan Magmatic Province, Avanzinelli et al., 2009; Conticelli et al., 2010, 2015a). Legend: 1) sedimentary substrate; 2) domes and lava flows of the I Cimino phase; 3) pyroclastic flows of II Cimino phase; 4) latitic to trachitic lava flows of the III Cimino phase; 5) olivine latitic to shoshonitic lava flows of the IV Cimino phase; 6) Roman-type Leucite -bearing Vico volcanic rocks (Roman Magmatic Province); 7) quaternary fluvial deposits; (redrawn after Perini et al., 1997, 2004; Conticelli et al., 2007, 2010; 2013, 2015a). (b) Geological sketch map the Vico volcano. Legend: E) caldera rim; 8) Macigno Flysch; 9) Allochtonous Flysch; 10) post-orogeny Pliocene-Pleistocene marine sediments; 11-13) Cimino dome complex (1300-900 ka): 11) lamproitic lava; 12) rhyodacitic domes; 13) rhyodacitic Peperino Tipico ignimbrite; 16-14) Lago di Vico synthem (second period; the earlier Rio Ferreira Synthem is not representable at this map scale): 14) stratocone building leucite-bearing lavas; 15) caldera forming phonolitic ignimbrites; 16) Carbognano phreatomagmatic ignimbrite; 17) Monte Venere lithosome: post-caldera scoria cones (third period); 18) travertine; 19) Holocene alluvial and lacustrine deposits; (c) Geological sketch section through the Monte Cimino (1.3-0.9 Ma) and Vico (0.45-0.05 Ma) nested volcanoes (redrawn after Conticelli et al., 2013; Avanzinelli et al., 2017).

(Nappi et al., 1994, 1995) and the interbedded ca. 332 ka Orvieto-Bagnoregio ignimbrite related to a major caldera-forming event (Palladino et al., 2014; Marra et al., 2019; Palladino & Pettini, 2020);

4) the Montefiascone lithosome, SE of Lake Bolsena, consists of a central volcanic edifice cut by a nearly circular polygenetic caldera, ca. 3 km across. The onset of activity (<286 ka), broadly contemporaneous with Latera, is represented by the lithic-rich Montefiascone basal ignimbrite (Nappi & Marini, 1986), a hydromagmatic event related to an early stage of caldera collapse;
5) Latera is a gently sloping, central volcanic edifice (at least 30 km across) located in the western part of Vulsini (Fig. 9). Its climactic activity (ca. 253-177 ka; Monaco et al., 2022a and reference therein) produced widespread Plinian fall (Palladino & Agosta, 1997) and ash-pumice flow deposits (Sparks, 1975; Palladino & Valentine, 1995; Palladino & Giordano, 2019; Valentine et al., 2019), resulting in the Latera-Vepe nested caldera system (9 km by 7 km across) through multiple collapse stages (e.g., Nappi, 1969; Sparks, 1975; Barberi et al., 1984; Vezzoli et al., 1987; Nappi et al., 1991; Palladino & Simei, 2005). In particular, the Sovana and Onano eruptions, which emplaced typical lag breccia deposits, are identified as major caldera-forming events (Palladino & Simei, 2005; Palladino et al., 2014). The late volcanic activity was characterized by effusive (e.g.,

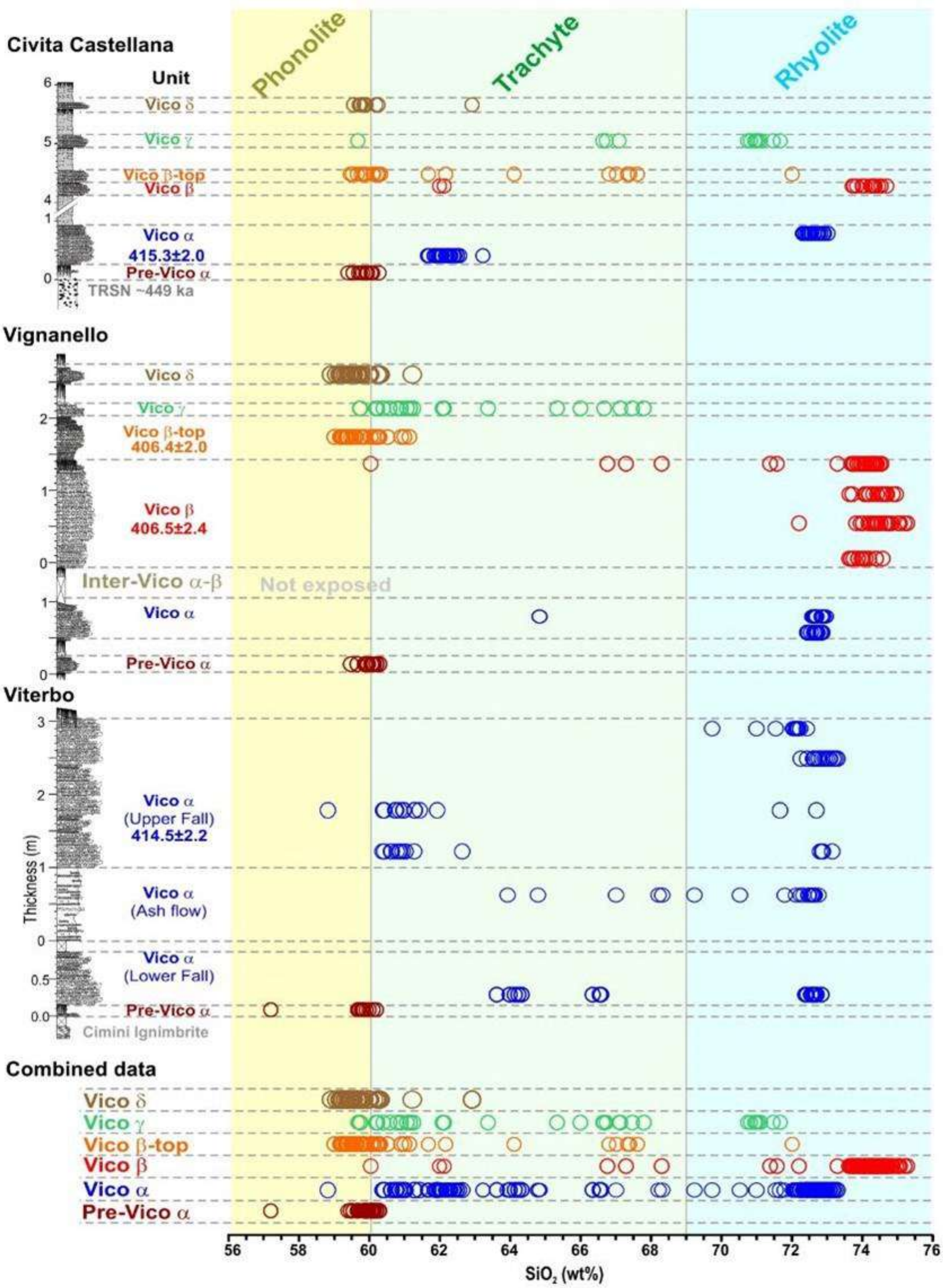


Fig. 11 -Variation of the silica content upsection within the Vico α and Vico β Plinian deposits and the minor pre-Vico α and post-Vico β units (modified from Pereira et al., 2020).

the ca. 157 ka shoshonitic Selva del Lamone lava plateau), Strombolian and phreatomagmatic eruptions from intra- and peri-caldera monogenetic centers.

The Vulsini volcanics span a full range of potassic rock types, from the least differentiated (i.e., K-basalts, trachybasalts, basanites, and tephrites), which mostly characterize lava flows and plateaus, as well as Strombolian and hydromagmatic products, to the most differentiated ones (i.e., phonolites, and trachytes), typical of Plinian pumice-fall and ash-pumice flow deposits from major explosive eruptions (Conticelli et al., 1987, 1991; Nappi et al., 1994; Palladino & Agosta, 1997; Palladino et al., 2010, 2014 and reference therein). In particular, the Montefiascone lavas (e.g., from the circum-caldera Orto Piatto eruptive center; ca. 226 ka; Nappi et al., 1995) display peculiar compositional features (i.e., high CaO and MgO and low SiO_2, Al_2O_3 and alkali contents) and are recognized among the most primitive rock types in the Vulsini district and in the whole Roman Province as well (Conte et al., 2009).

In distal setting, the activity of the Vulsini Volcanic District is particularly well documented for the Latera stage (ca. 255-157 ka) in Fucino Basin, central Italy (Monaco et al., 2021). From the same basin, several tephra layers originated from Vulsini have been also found in the interval 430-300 ka (e.g., Orvieto-Bagnoregio unit of 335 ka; Leicher et al., 2023; Monaco et al., 2021), while activity of the Paleovulsini is possibly documented in Lake Ohrid (Leicher et al., 2019; 2021).

3.2.3. Cimino (ca. 1.1-0.8 Ma) and Vico (ca. 415-95 ka)

On the basis of morphological, geological, and mineralogical data the Cimino and Vico volcanoes (Fig. 10) represent two distinct edifices active in different times, but fed by the same plumbing system (e.g., Perini, 1997; Perini et al., 1997, 2000, 2003, 2004; Peccerillo, 2005; Conticelli et al., 2010, 2013, 2015a;). According to Peccerillo et al. (1987) the Cimino volcanic complex, active during the Early Pleistocene, belongs to the Tuscan Magmatic Province, whilst the nested and younger Vico stratovolcano, active during the Late Pleistocene, belongs to the Roman Magmatic Province.

The Cimino volcanic complex is the oldest edifice of the Cimino-Vico area, and it is a relatively simple volcano with an evolution through four main phases (Conticelli et al., 2013) in a short time span from 1.35±0.08 to 0.94±0.20 Ma (e.g., Evernden & Curtis, 1965; Nicoletti, 1969; Borghetti et al., 1981).

The Cimino volcano is made up of: i) an initial alignment of trachydacitic to trachytic cryptodomes along a NW-SE trend (auctor. “peperino delle alture”); ii) two welded pyroclastic flow units, outpoured when the activity was concentrated in the central part of the volcano (auctor. “peperino tipico”); iii) lava flows of the Monte Cimino cone; iv) final mafic lava flows unit (e.g., Puxeddu, 1971; Lardini & Nappi, 1987; Perini et al., 2003; Cimarelli & de Rita, 2006a,b; LaBerge et al., 2006, 2008; Conticelli et al., 2013; Nappi et al., 2022).

The Cimino volcanic rocks are invariably leucite-free ranging with time from trachydacite, to trachyte, latite, olivine-latite, and shoshonites. Details on mineralogical and petrographic features are provided in Perini et al. (2003) and Conticelli et al. (2013).

After a hiatus of some hundreds of thousands of years, renewal of volcanic activity occurred in the area, at 450 ka, few kilometers southwest of the Monte Cimino with eruption of Roman-type, silica-under-saturated, leucite-bearing magmas (Fig. 10). The Vico volcano was built in the form of a conic stratovolcano cut by a summit polygenetic caldera (Perini et al., 1997, 2004). Post-caldera activity, subordinate in volume, both within and on the N and W edges of the caldera do occur (Fig. 10b; Perini & Conticelli, 2002; Perini et al. 1997, 2004). Vico volcanic rocks were strong to mild silica-undersaturated, leucite- to plagioclase-leucite-bearing generated by partial melting of a lithospheric mantle wedge under high X_{CO2} conditions.

The eruptive history of the Vico volcano is subdivided into three periods and corresponding lithosomes (Perini et al., 2004; Conticelli et al., 2013):

1) Rio Ferriera Synthem (sensu Perini et al., 2004), involving an early effusive period, alternating with explosive phases at ca. 415-400 ka that produced the Vico α and Vico β Plinian fall markers (Cioni et al., 1987; Pereira et al., 2020);
2) Lago di Vico Synthem, involving the build-up of the main stratovolcano by essentially effusive activity (ca. 305-258 ka), followed by a climactic, caldera-forming, explosive phase at ca. 250-144 ka that produced ignimbrites A, B, C (also known as Tufo Rosso a Scorie Nere Vicano) and D (Locardi, 1965; Bertagnini & Sbrana, 1986), corresponding to the Farine, Ronciglione, Sutri and Carbognano formations (Perini et al., 2004; Bear et al., 2009a,b);
3) Monte Venere Synthem (ca. 138-95 ka), characterized by phreatomagmatic and Strombolian-effusive activities from peri- and intra-caldera monogenetic centers.

Vico α and Vico β are geochemically peculiar within the peri-Tyrrhenian volcanism. In fact, in addition to the usual trachy-phonolitic ones, the juvenile fraction of the deposits of both eruptions contains rhyolitic pumice clasts (Fig. 11), which represent a puzzling exception among the Quaternary potassic volcanoes of central Italy. Isotopic and trace element data indicate an origin of these rhyolitic magmas by fractional crystallization processes (Perini et al., 2004), occurred at unusually high-pressure conditions: the rhyolitic glasses in pumice clasts show, indeed, high volatile contents (H_2O ca. 5-7 wt%) and are in equilibrium with amphibole (Monaco, 2022). The deep level of the Vico α and Vico β pre-eruptive magma systems is consistent with the wide dispersal of the fallout products and the lack of co-eruptive pyroclastic currents associated with caldera-forming processes.

Conversely, the Tufo Rosso a Scorie Nere Vicano (ca. 150 ka; Laurenzi & Villa, 1987) is related to the main caldera collapse at Vico (Bear et al., 2009a,b; Palladino et al., 2014). The eruption succession includes: i) a basal Plinian fall horizon containing whitish sub-aphyric pumice, related to a central feeder conduit; ii) black spatter-bearing, coarse lithic-rich, co-ignimbrite lag breccias, related to developing ring faults during caldera collapse; iii) red tuff with leucite-bearing black scoria, that is the main pyroclastic flow deposit emplaced at the eruption climax, extending with remarkable continuity as far as 25 km from the vent, over an area of 1200 km^2. The facies architecture and the variation patterns of juvenile textural features and glass compositions, along with mineral chemistry, reveal a magma chamber characterized by thermal and volatiles gradients (which reflect in crystal concentration gradients) and changing magma withdrawal dynamics in the course of a typical caldera-forming event (Bear et al. 2009a; Palladino et al., 2014).

Owing to their peculiar rhyolitic components and/or wide dispersal, Vico α and Vico β and the Vico ignimbrites C and B are key stratigraphic markers for regional and ultra-regional correlations (Cioni et al., 1987; Pereira et al., 2020; Monaco et al., 2021; Iurino et al., 2022), as

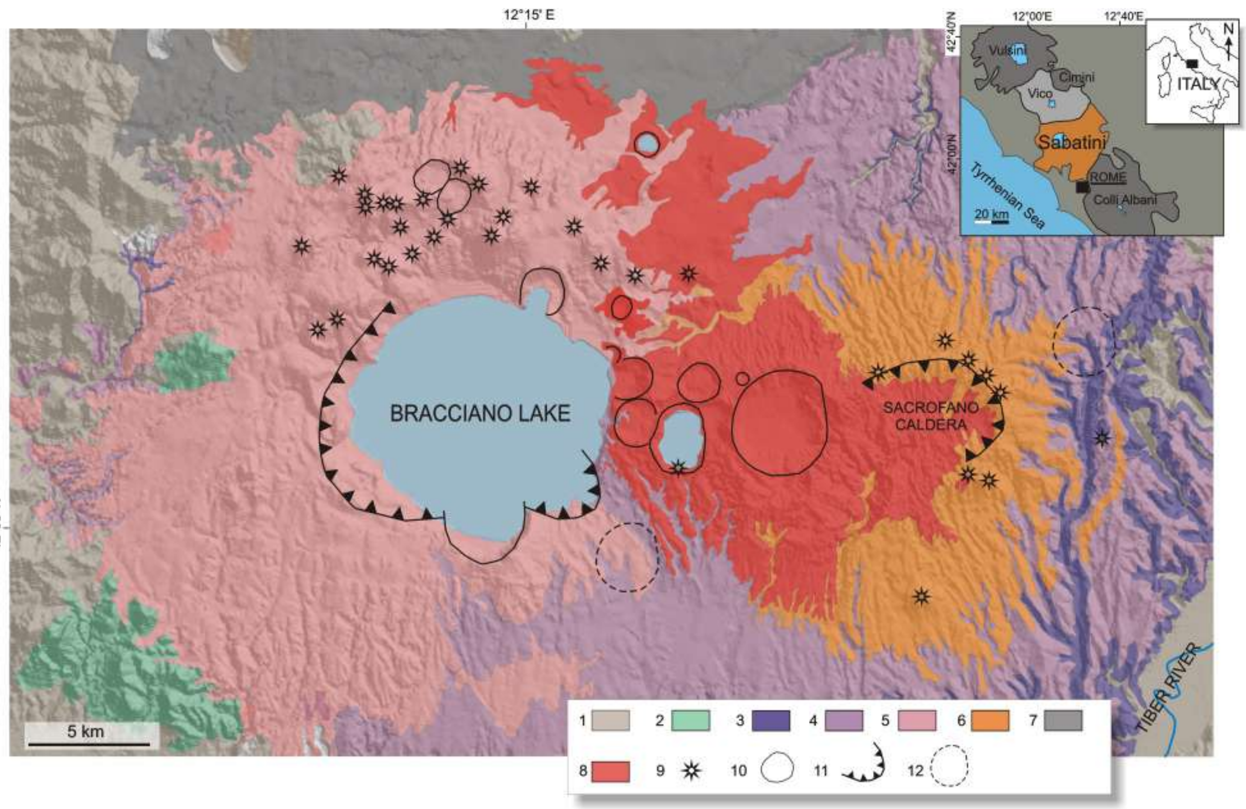


Fig. 12 - Sketch geological map of the Sabatini Volcanic District (after Sottili et al., 2010a), showing the areal distributions of the main activity periods. Insets show the locations of the Sabatini and the other volcanic districts of the Roman province, central Italy. Legend: 1- Sedimentary terrains; 2-M.ti Ceriti-Tolfetano-Manziate lava domes (Pliocene); 3- Morlupo phase products (0.6-0.5 Ma); 4- Southern Sabatini products (0.5-0.4 Ma); 5- Bracciano Caldera products (0.3-0.2 Ma); 6- Sacrofano Caldera products (0.3-0.2 Ma); 7- Vico Volcano (0.4-0.1 Ma); 8- Sabatini Late activity (< 0.15 Ma); 9- Scoria cones; 10- Maars; 11- Caldera rims; 12- Buried source areas.

far as Lake Ohrid, Albania-North Macedonia (Leicher et al., 2021).

3.2.4. Sabatini Volcanic District (ca. 810-70 ka)

The Sabatini Volcanic District extends over an area of ca. 1,800 km^2 north of Rome (Fig. 12). Volcanism was dominated by explosive activity ranging through a broad spectrum of intensities and magnitudes, from Strombolian and phreatomagmatic eruptions from monogenetic centers to Plinian and pyroclastic flow-forming events associated with caldera collapses (Sottili et al., 2004, 2010a; Marra et al., 2014, 2016). Today, the main landforms in the area are represented by the Bracciano caldera, hosting a lake ca. 9 km in diameter, and by the 6 x 4 km Sacrofano caldera to the East. Although quite subordinate volumetrically, a number of monogenetic tuff rings and scoria cones that postdate the main caldera-forming events are either clustered or scattered in the northern area of the district and between the Bracciano and Sacrofano calderas (Sottili et al., 2010a; Valentine et al., 2015).

Eruptive activity mostly took place in the 0.6-0.1 Ma time span, although older widespread Plinian fall deposits of uncertain source (yet tentatively attributable to a Sabatini paleo-activity) were also found in drill sites along the Tiber Valley and dated at ca. 810-614 ka (Karner et al., 2001). Conventionally, the Sabatini activity is divided into three main phases (Cioni et al., 1987, 1993; de Rita et al., 1993, 1996; Barberi et al., 1994; Conticelli et al., 1997; Karner et al., 2001; Sottili et al., 2004, 2010, 2012, 2019; Masotta et al., 2010):

1) early phase from the Morlupo (ca. 589-500 ka) and Southern Sabatini (ca. 500-400 ka) source areas, respectively to the East and South of the present Lake Bracciano, characterized by widespread pyroclastic flows and Subplinian to Plinian fallout (VEI 3-5). During this phase of activity, the main eruptions of the ca. 548 ka Tufo Giallo della Via Tiberina and the ca. 449 ka Tufo Rosso a Scorie Nere Sabatino occurred, the latter being the volumetrically most important, caldera-forming event in the whole activity history of Sabatini (Sottili et al., 2004; 2010; Masotta et al., 2010; Marra et al., 2014; Palladino et al., 2014);
2) intermediate phase of effusive and explosive activity (ca. 400-200 ka) from Lake Bracciano and Sacrofano source areas, marked by two major caldera-forming eruptions, respectively the ca. 310 ka Tufo di Bracciano and the ca. 285 ka Tufo Giallo di Sacrofano;
3) late eruptive phase (ca. 150-70 ka), dominated by phreatomagmatic and subordinate Strombolian and effusive activities from monogenetic and polygenetic tuff rings and scoria cones in the central area of Sabatini (Sottili et al., 2012; Valentine et al., 2015), mostly between the Bracciano and Sacrofano calderas (e.g., Baccano, Martignano and Stracciacappa centers).

Compositionally, the Sabatini volcanics cover a wide spectrum of potassic rock types, ranging from trachybasalts to trachytes and phonolites (Conticelli et al., 1997, 2010; Peccerillo, 2005; Sottili et al., 2019). Marra et al. (2020b) highlighted a noticeable parallelism in time and eruption magnitude between Sabatini and Colli Albani during most of their lifetime, and suggested a common regional tectonic trigger, such to define them twin volcanic systems.

The explosive activity of the Sabatini Volcanic District is well documented in distal setting. The most widespread unit is Fall A (~500 ka) of the Morlupo phase, found in central-southern Italy intermountain basins (Sulmona, Mercure and Acerno; Giaccio et al., 2014; Petrosino et al., 2014; Di Rita & Sottili, 2019) and Lake Ohrid (Leicher et al., 2021). Many other tephra layers spanning the interval 550-120 ka, of either well identified or supposed Sabatini origin, have been found in Fucino and Acerno basins, Lake Monticchio and even Lake Ohrid (Wulf et al., 2012; Giaccio et al., 2012; 2017a, 2019; Petrosino et al., 2014; Monaco et al., 2021; 2022a; Leicher et al., 2023). Tephra layers with Sr isotope composition and age comparable to the putative Sabatini paleo-activity (~615-810 ka) have been found in the Sulmona basin (Giaccio et al., 2013b; 2015; Sottili et al., 2019), although their attribution to the Sabatini Volcanic District is still uncertain.

3.2.5. Colli Albani Volcanic District (ca. 608-36 ka)

Following a poorly documented early activity period, the Colli Albani volcanism took place in three main phases marked by different eruptive mechanisms and magma volumes (de Rita et al., 1988, 1995b; Giordano et al., 2006; Fig. 13):

1) the climactic activity period, known as Tuscolano-Artemisio Phase (de Rita et al., 1988) or, referring to the evolution of the edifice, as Vulcano Laziale caldera complex (Giordano & the CARG Team, 2010), produced at least five subsequent eruptive cycles, dominated by large mafic and caldera-forming ignimbrites, in the ca. 561-351 ka time span (Karner et al., 2001; Marra et al., 2009), alternated with effusive and mild explosive activity. Strikingly, large-volume (up to several tens of km^3) pyroclastic flows were fed by K-foiditic magmas with SiO_2 as low as 42 wt%, e.g., during the early Trigoria-Tor de' Cenci and Palatino "pisolitic tuff" eruptions and the main ca. 456 ka Pozzolane Rosse and ca. 365 ka Villa Senni caldera-forming eruptions (Freda et al., 1997, 2011; Palladino et al., 2001; Boari et al. 2009a; Marra et al., 2009; Vinkler et al., 2012; Gaeta et al., 2016);
2) the Faete Phase (de Rita et al., 1988), during which the Tuscolano-Artemisio peri-caldera ring fracture system and the intra-caldera Faete stratovolcano were built (Giordano & the CARG Team, 2010), occurred at ca. 308-250 ka (Marra et al., 2003) and was characterized mainly by Strombolian and effusive activities;
3) following a ca. 50 kyr-long dormancy, the Late Hydromagmatic Phase (ca. 200-36 ka; Marra et al., 2003, 2016; Freda et al., 2006; Giaccio et al., 2007, 2009), or Via dei Laghi lithosome (Giordano & the CARG Team, 2010), was dominated by maar and tuff ring forming eruptions (e.g., Ariccia, Nemi, Valle Marciana, Laghetto or Giuturna, Albano). In particular, the Albano polygenetic maar (ca. 69-36 ka) hosted the most recent and energetic activity (Giordano et al. 2002; Freda et al., 2006; Giaccio et al., 2007; De Benedetti et al., 2008).

The age of the last eruption at Colli Albani has been a matter of debate (see the above references and Soligo & Tuccimei, 2010 for a review). A wide data-set of accurate $^{40}Ar/^{39}Ar$ dating consistently indicates that the last documented eruptive event occurred at the Albano maar at ca. 36 ka (Marra et al., 2003; Giaccio et al., 2007, 2009), in spite of younger ages reported in the literature (e.g., ca. 26 ka; Villa et al., 1999) that are affected by stratigraphic inconsistencies and/or methodological uncertainties. In addition, a series of lahar deposits (Tavolato Formation) have been attributed to the repeated overflows of the Lake Albano and possibly associated with intrusion-triggered limnic eruptions (Funiciello et al., 2003; De Benedetti et al., 2008), although the occurrence and exact age of these events, possibly extending into the Holocene, has been a matter of controversy (Marra & Karner, 2005; Giaccio et al., 2007, D'Ambrosio et al., 2010). On these grounds, bv considering the long-lasting quiescence periods occurred in the past (ca. 40 kyr; Marra et al. 2004), it cannot be excluded that Colli Albani can be potentially active (de Rita et al., 1995a; Marra et al., 2016), also in light of the several geophysical (Chiarabba et al., 1997) and geochemical (Carapezza et al., 2010) indicators of volcanic unrest.

Although the Colli Albani rock-types span the tephrite, phonotephrite and K-foidite fields of the TAS classification diagram, they show invariably, regardless of eruption size, a leucititic mineral assemblage. The latter is characterized by the virtual absence of feldspars, with a few exceptions in the groundmass of olivine-bearing lavas, and in glass-bearing juvenile pyroclasts formed at high cooling rates (Trigila et al., 1995; Gaeta, 1998; Gaeta & Freda, 2001; Gaeta et al., 2006, 2016; Giordano et al., 2006; Boari et al., 2009a; Marra et al., 2009; Gozzi et al., 2014). The peculiar magma differentiation trend, leading to K-foiditic magmas with SiO_2 contents <50 wt%, has been explained by the assimilation of sedimentary carbonate. This process allows the establishment of CO_2-rich conditions in the Colli Albani magmas and reduces the stability fields of olivine and phlogopite in favour of clinopyroxene and leucite (Gaeta et al., 2000, 2009, 2021; Freda et al., 2008; Di Rocco et al., 2012). The anomalous highly explosive behaviour of SiO_2-poor magmas at Colli Albani has also been attributed to CO_2 entrainment in the pre-eruptive system due to magma-carbonate wall rock interaction at supra-crustal levels (Freda et al., 1997, 2011; Sottili et al., 2010b; Cross et al., 2014). An alternative hypothesis suggests that the main location for magma-carbonate fluids interaction, controlling both the peculiar line of descent and the explosivity, is within the metasomatized mantle source (Boari et al., 2009a; Jorgenson et al., 2020).

The uncommon lithology and K-foiditic composition and other peculiar features of the Colli Albani tephra (e.g., the $^{87}Sr/^{86}Sr$ time-dependent variability; Gaeta et al., 2016) make them among the most distinctive markers in the framework of the central Mediterranean tephrostratigraphy (Giaccio et al., 2013a). The pyroclastic products of the main activity phase of Colli Albani (ca. 561-351 ka), as well as of the Late Hydromagmatic Phase (ca. 200-36 ka), are widely dispersed in central-southern Italy (Giaccio et al., 2009; 2013; 2017a; 2019; Petrosino et al.,

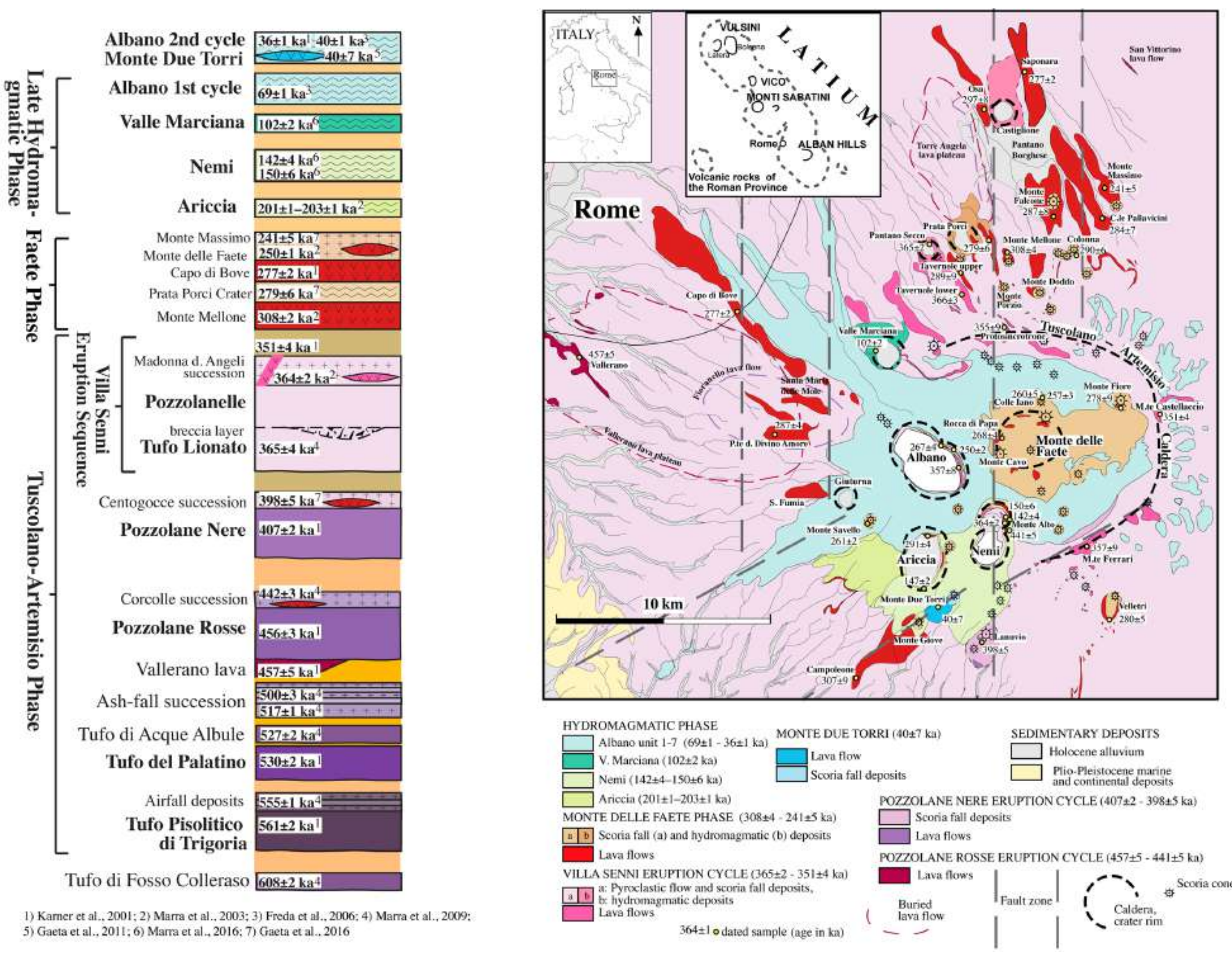


Fig. 13 - Stratigraphic-chronological scheme and geological sketch of the Colli Albani (Alban Hills) Volcanic District (Marra et al., 2011; Gaeta et al., 2016).

2014; Monaco et al., 2021). The largest events, such as the ~456 ka Pozzolane Rosse, can be traced in the tephra record as far as Lake Ohrid (Leicher et al., 2019; 2021).

3.2.6. Volsci Volcanic Field (ca. 760-230? ka)

This volcanic field (also known as Monti Ernici or Middle Latin Valley volcanoes) groups several monogenetic centers (e.g., scoria cones, maar-diatremes and tuff rings) scattered through the Middle Latin Valley and the Volsci carbonate range, as far as the adjoining Pontina Plain (Pasquarè et al., 1985; Boari et al., 2009b; Centamore et al., 2010; Cardello et al., 2020; Marra et al., 2021 and reference therein; Fig. 14). The Volsci eruptive pattern was characterized by small volume (in the order of 0.01-0.1 km^3) eruptions from a network of monogenetic centers, totaling nearly 4 km^3, i.e., two orders of magnitude lower than the major potassic volcanic districts of the Roman Province. Three main phases of activity have been distinguished (Marra et al., 2021), as follows:

1) the "Early Eruptive Phase" (ca. 761-541 ka) mainly occurred along the carbonate Volsci Range (e.g., Patrica eruptive center) and was essentially fed by K-rich, leucite-melilite-bearing (HKS) magmas, showing evidence of high degree of carbonate assimilation, possibly due to multi-stage ascent through thick carbonate successions (hence the diffuse phreatomagmatic character);
2) subsequently, during the "Main Eruptive Phase" (ca. 424-349 ka), HKS rock types alternated with poorly differentiated, plagioclase-bearing (KS) rock types from eruptive sources close in time and space (e.g., in the Ceccano and Pofi areas). In particular, the eruptive activity along the major faults bounding the Middle Latin Valley Graben testifies for the fast ascent of nearly primary KS magmas (K-basaltic in composition; Centamore et al., 2010);
3) during the "Late Eruptive Phase (ca. 300-231? ka), KS magmas prevailed, although still accompanied by HKS events. The low-intensity and low-volume (i.e., Hawaiian-Strombolian, phreatomagmatic, and subordinate effusive) eruptive events, fed by poorly differentiated potassic magmas, reflect the tectonically controlled, fast ascent of primitive magma batches from the mantle source ("bullet eruptions"; Cardello et al., 2020).

Owing to the lack of glass chemical compositions from the Volsci pyroclastic deposits, identifying potential distal correlatives is hardly feasible. Nevertheless, the

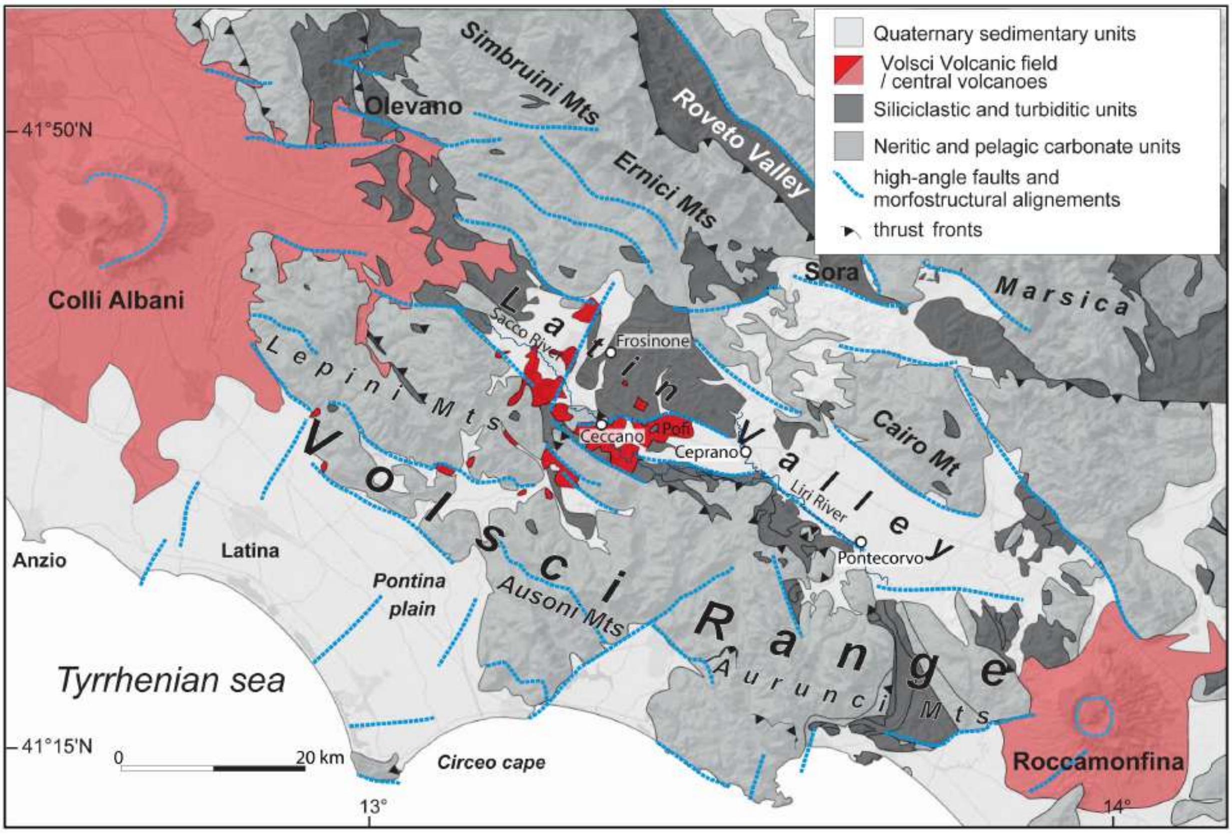


Fig. 14 - Geological sketch map of southern Latium, showing the location of the Volsci Volcanic Field (square area) between the major Colli Albani and Roccamonfina volcanoes (after Marra et al., 2021, modified).

Volsci is currently the sole volcanic source for which an activity as old as 760 ka is documented in near-vent setting, which is geochronologically compatible with a series of Sulmona tephra layers dated between ~805 and ~715 ka (Giaccio et al., 2013b; 2015).

3.2.7. Roccamonfina Volcano (630-55? ka)

Roccamonfina volcano (Fig. 15) is the result of the superposition of different morphostructures. The oldest edifice is a stratovolcano made up mostly of leucite-bearing, highly silica-undersaturated (High-K series; HKS; Conticelli et al., 2009b), phonotephritic lava flows and Strombolian deposits (Roccamonfina Synthem, de Rita & Giordano, 1996). An age of 630 ka was obtained for the pyroclastic deposits at the very base of the volcanic pile sampled from a deep borehole drilled at the center of the present caldera (Ballini et al., 1989a). Some K-Ar ages as old as ~1.2 Ma (Gasparini & Adams, 1967) and ~700 ka (Radicati di Brozolo et al., 1988) reported for the products of this early stage may be not reliable for modern analytical standards. As a result of NE-trending extensional faults cutting the edifice, a major lateral sector collapse interrupted the growth of the stratovolcano (de Rita & Giordano, 1996) at around ~440 ka (end of Rio Rava stage; Chiesa et al., 1995; Giannetti, 2001; Rouchon et al., 2008).

Subsequent activity (related to the Riardo Synthem; de Rita & Giordano, 1996) shifted to dominantly explosive and a series of VEI 3-5 Subplinian to Plinian eruptions shaped a polygenetic summit caldera, superimposed to the earlier sector collapse (Cole et al., 1993; de Rita & Giordano, 1996; Giannetti, 2001). Ruchon et al. (2008) subdivided this caldera-forming stage into several substages. The oldest one, between 410 ka and 360 ka, culminated with the first caldera-forming event of the Brown Leucitic Tuff (BLT; Luhr & Giannetti, 1987; Cole et al., 1993), which is the last eruption involving HKS magmas at Roccamonfina. The following activity was fed by slightly silica-saturated leucite-free magmas with a shoshonite affinity (SHO; Conticelli et al., 2009b). Between ~331 ka and ~230 ka, the White Trachytic Tuffs (WTTs; Ballini et al., 1989b; Cole et al., 1993; Giordano, 1998a,b: Giannetti & De Casa, 2000) originated from multiple trachytic Plinian eruptions. At least two of them (i.e., WTT Cupa and WTT Galluccio of Giordano, 1998a, or LWTT and UWTT of Giannetti & De Casa, 2000) resulted into caldera collapses in the central area of the volcano. The pyroclastic deposits of the Yellow Trachytic Tuff (Rouchon et al., 2008) were also emplaced during this period.

The youngest activity at Roccamonfina (Vezzara Synthem; de Rita & Giordano, 1996; or 170-150 ka Santa Croce stage; Rouchon et al., 2008), was controlled by N-trending tectonic lineaments and gave birth to the intra-caldera latitic domes of Lattani and M.te Santa Croce. Eruptive activity ended at ~150 ka according to Rouchon

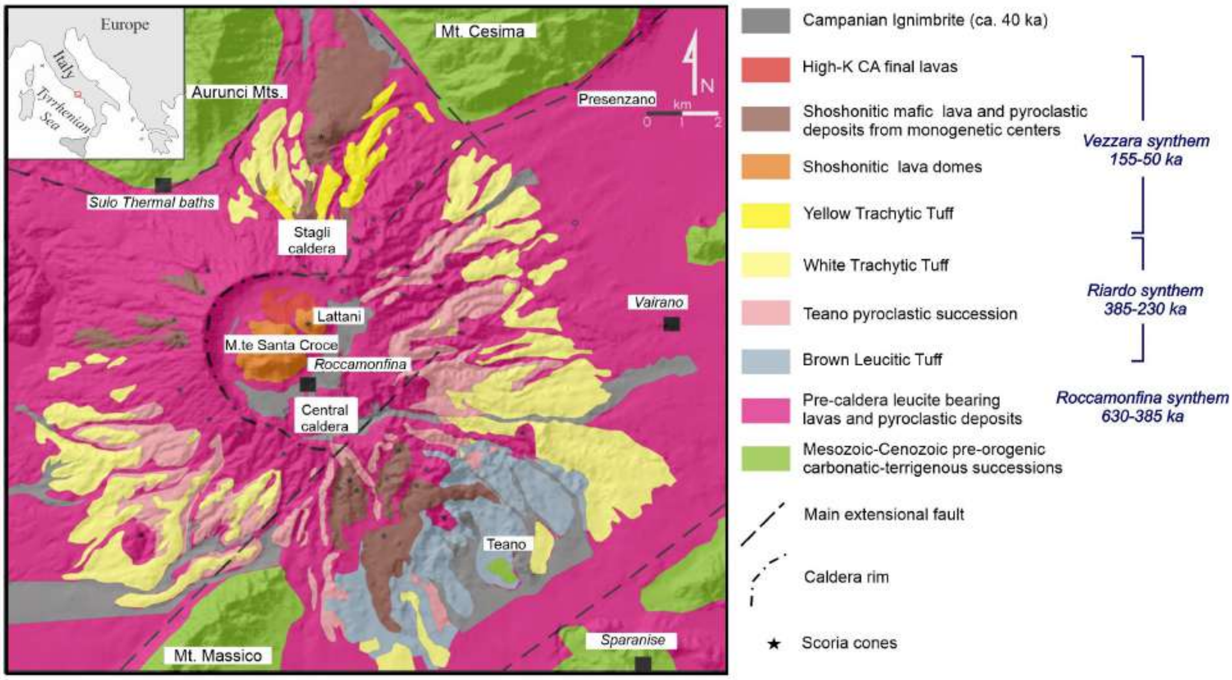


Fig. 15 - Sketch geological map of the Roccamonfina Volcano showing the main volcanic units grouped into the respective synthems (redrawn after Conticelli et al., 2010).

et al. (2008), though a younger age of ~55 ka is reported in Radicati di Brozolo et al. (1988).

In distal setting, tephra layers attributed to the Roccamonfina activity have been recognized on regional and extra-regional scales. Rio Rava is documented in Molise region, immediately east of Roccamonfina volcano (Russo Ermolli et al., 2010), while BLT has been recently identified in the Fucino basin (central Italy) and associated to the widespread "Bag tephra" of Danubian Middle Pleistocene loess succession (Leicher et al., 2023). The LWTT is also well documented in Molise alluvial-lacustrine successions (Amato et al., 2014; Galli et al., 2017) and, more distally, in the Fucino basin (Leicher et al., 2023). Of note, distal records, spanning the 600-200 ka interval, indicate the occurrence of a number of tephra layers with Roccamonfina-like geochemical compositions that currently lack specific proximal counterparts (e.g., Giaccio et al., 2014; Petrosino et al., 2014; Leicher et al., 2021), providing evidence for a conspicuous, yet still not well-defined, Middle Pleistocene explosive activity at Roccamonfina volcano.

3.2.8. Final remarks: the state of volcanic activity

Ongoing research aims at refining the chronology of volcanic activity in central Italy, in order to better constrain the eruption frequency and the duration of quiescence periods, which is crucial for long-term hazard assessment. In this regard, tephra correlations from distal archives provide additional, detailed information for a comprehensive reconstruction of tempo and dynamics of the peri-Tyrrhenian Quaternary volcanoes (e.g., Giaccio et al., 2019; Leicher et al., 2019; Monaco et al., 2021; 2022a).

Based on the ages of the last documented eruptive events in the Roman Province, i.e., ca. 111 ka at Vulsini, ca. 95 ka at Vico, ca. 70 ka at Sabatini and 36 ka at Colli Albani (the latest activity at Volsci and Roccamonfina is not yet well constrained), it appears that the waning (or, possibly, cessation) of eruptive activity shifted progressively from NW to SE. Strikingly, these volcanoes reveal a "relay behavior" (Marra et al., 2020a), i.e.: the waning stage of activity at Vulsini (ca. 150-110 ka) was concomitant to an activity climax at Vico (third period, Perini et al., 2004; also including the main caldera-forming event of Tufo Rosso a Scorie Nere Vicano); then, waning activity at Vico (ca. 95 ka) occurred during a peak of phreatomagmatic activity at Sabatini; again, the last documented eruption at Sabatini (ca. 70 ka) was accompanied by the onset of the Albano maar cycle at Colli Albani. Finally, the last eruptive phase at Albano (40-36 ka) was broadly coincident with the climactic super-eruption of the 40 ka Campanian Ignimbrite at Campi Flegrei. The focus of volcanic activity seems therefore to have shifted through time from Northern Latium, through the Roman area, toward the Neapolitan area. However, considering that the time lapse since the last documented eruptions at Sabatini (ca. 70 kyr) and Colli Albani (ca. 36 kyr), is in the same order of the dormancies occurred in the activity history of the two volcanoes (ca. 70 kyr and ca. 40 kyr, respectively), the possible occurrence in the near future of eruptive codas from the dormant "twin" volcanoes at the gates of Rome cannot be ruled out (Marra et al., 2020b).

In addition, a recent work by Giordano & Caricchi (2022) points out that defining the state of activity of volcanoes based on recurrence times alone requires the assumption that the magma plumbing system is steady-state, while even at constant magma fluxes, the thermal and geometric configurations and chemical reactivity of plumbing systems may change significantly over time. This means that as they thermally mature, plumbing systems may allow the growth and coalescence of progressively larger magma bodies, implying progressively a longer quiescence, which may lead to larger eruptions. In this sense, refining our knowledge of the longevity of a volcano is as important as knowing the age of its last eruption, and this is why recent efforts in extending back the distal tephrochronology of the peri-Tyrrhenian Quaternary volcanoes is so important (e.g., Giaccio et al., 2019; Leicher et al., 2019; Monaco et al., 2021). At the same time Giordano & Caricchi (2022) pointed out that the heat flux of the Roman volcanoes, especially at Vulsini and Sabatini is as high as, and locally even higher than, at the certainly active Ischia and Campi Flegrei volcanoes and much higher than at Vesuvius, suggesting that in order to define the state of activity of these volcanoes a much better imaging of their present magma plumbing system is required, although again none should be considered extinct.

4. PENINSULAR-INSULAR SOUTHERN ITALY VOLCANOES AND SEAMOUNTS

4.1. The Pontine Archipelago (A.M.C., C.P.)

The Pontine Archipelago consists of five major volcanic islands divided in western islands, Ponza, Palmarola and Zannone and eastern Islands, Ventotene and S. Stefano (Fig. 16a). The western islands lie on a NE-SW elongated structural high dividing two major areas of sedimentation i.e., the Palmarola and Ventotene intraslope basins (Fig. 16a) created by the Plio-Pleistocene extensional deformations, which also gave rise to a very steep NW-SE trending continental slope and an intense magmatic activity developed from late Pliocene to Late Pleistocene (Conte & Dolfi, 2002; Cadoux et al., 2005). The eastern islands are the emerged portions of the caldera rim of a large strato-volcano rising ca. 700 m from the sea-floor emplaced at the center of the subsiding Ventotene basin (Fig. 16a; Metrich et al., 1988; Bellucci et al., 1999).

In the western islands the Pliocene volcanic cycle (4.5-2.9 Ma; Cadoux et al., 2005; Fig. 17a) produced a large effusion of High-K, Calc-Alkaline (HKCA) rhyolite lavas from extensional fissures in a submarine environment. This led to the emplacement of differently textured hyaloclastites and lava dykes which formed the typical cryptodome-dyke systems that constitute most part of Ponza Island (Fig. 16b, c).

Indeed, three main coalescing domes of about 1 km radius, centered at Cala M.te Pagliaro and aligned along a NE trending regional fracture, have been identified at Ponza on a morphological basis (de Rita et al., 2001). The same likely occurred in the submerged SW sector of Palmarola (Conte et al., 2016; Fig. 16b). At Zannone, as a consequence of local eustatic movements, rhyolite magmas similar, and probably coeval to those of Ponza (Conte et al., 2016), were emplaced as cryptodomes in a subaqueous/subaerial environment and also intruded into a substrate made up of sedimentary and metamorphic units, which are locally exposed (de Rita et al., 1986). This prevented the fine white hyaloclastite to form and led to the formation of a strongly brecciated lava in which flow structures are often recognized. Similar brecciated lava facies are observed in the NW sector of Ponza island, including the islet of Gavi, suggesting a similar subaqueous/subaerial environment of emplacement.

The second volcanic cycle started in Early Pleistocene age (1.64-1.52 Ma; Fig. 17a) during which Palmarola was entirely built owing to the emplacement of a submarine hyaloclastite unit intruded by dykes ranging in compositions from near- to peralkaline trachytes and rhyolites belonging to the Transitional Rock-series (TR; Conte et al., 2016). The Pleistocene activity progressed with the local emission of comendite lava (1.2 Ma; Savelli, 1987) and the emplacement of pyroclastic trachytic products in the southern part of Ponza (1.2-0.9 Ma; Savelli, 1987; Bellucci et al., 1999; Fig. 17a), where the resumption of volcanism coincides with the transition from a submarine to a subaerial environment. The Pleistocene cycle ended at ca. 1.0 Ma with the emplacement of trachytes of potassic series (K-alkaline, KA Fig. 17b) in the southeastern part of Ponza (i.e., the Monte la Guardia lava dome, Punta della Guardia lava dyke, Fig. 16c). This episode represents the debut of the potassic alkaline magmatism that successively developed southeastward to the eastern Pontine islands up to the Roman-Campanian Magmatic Provinces. The eastern Pontine islands, Ventotene and S. Stefano, display a similar chronological sequence of the outcropping units, which were erupted during the 0.9 and 0.1 Ma time span (Fig. 17a). It was initially characterized by effusive activity comprising a number of trachybasaltic lava flows (Fig. 17b) followed by a huge explosive phase, which ended with the caldera formation. The composition of the volcaniclastic products ranges from latitic to trachytic and phonolitic (Fig. 17b; Bellucci et al., 1999).

Further insights on the magmatic spectrum existing in the Pontine Archipelago (Fig. 17b) were provided by submarine rocks of relatively undifferentiated compositions (basalt to andesites), so far missing in the literature, collected in outcrops along the continental slope (Conte et al., 2016), and in submarine volcanic edifices (i.e., the Ventotene Volcanic Ridge; Conte et al., 2020; Fig. 16a, 17a, 17b). Geochemical data from these new findings, and thermodynamic modeling allowed to rule out an anatectic origin for both subalkaline and peralkaline rhyolites recognized in the Archipelago (Paone, 2013) and to infer that the whole HKCA-and TR-series felsic rocks, and the KA as well, derived by the respectively poorly evolved rocks by fractional crystallization processes (Fig. 17b). Moreover, the new geochemical data confirm the orogenic signature of the three suites and indicate highly heterogeneous mantle sources, due to crustal components variously recycled in the mantle via subduction. In addition, the age of these new findings (Conte et al., 2020) led to infer that the Pliocene and Pleistocene volcanic cycles developed in the Pontine Archipelago are likely associated to the Pliocene E-W directed rifting stage of the Tyrrhenian back-arc basin, which produced a more intense activity during the time span ~5.0-2.0 Ma (the emplacement time of HKCA and TR series). After that, this fault activity decreased and the extensional direction changed to NW-SE causing the beginning of the KA activity (<~2.0 Ma).

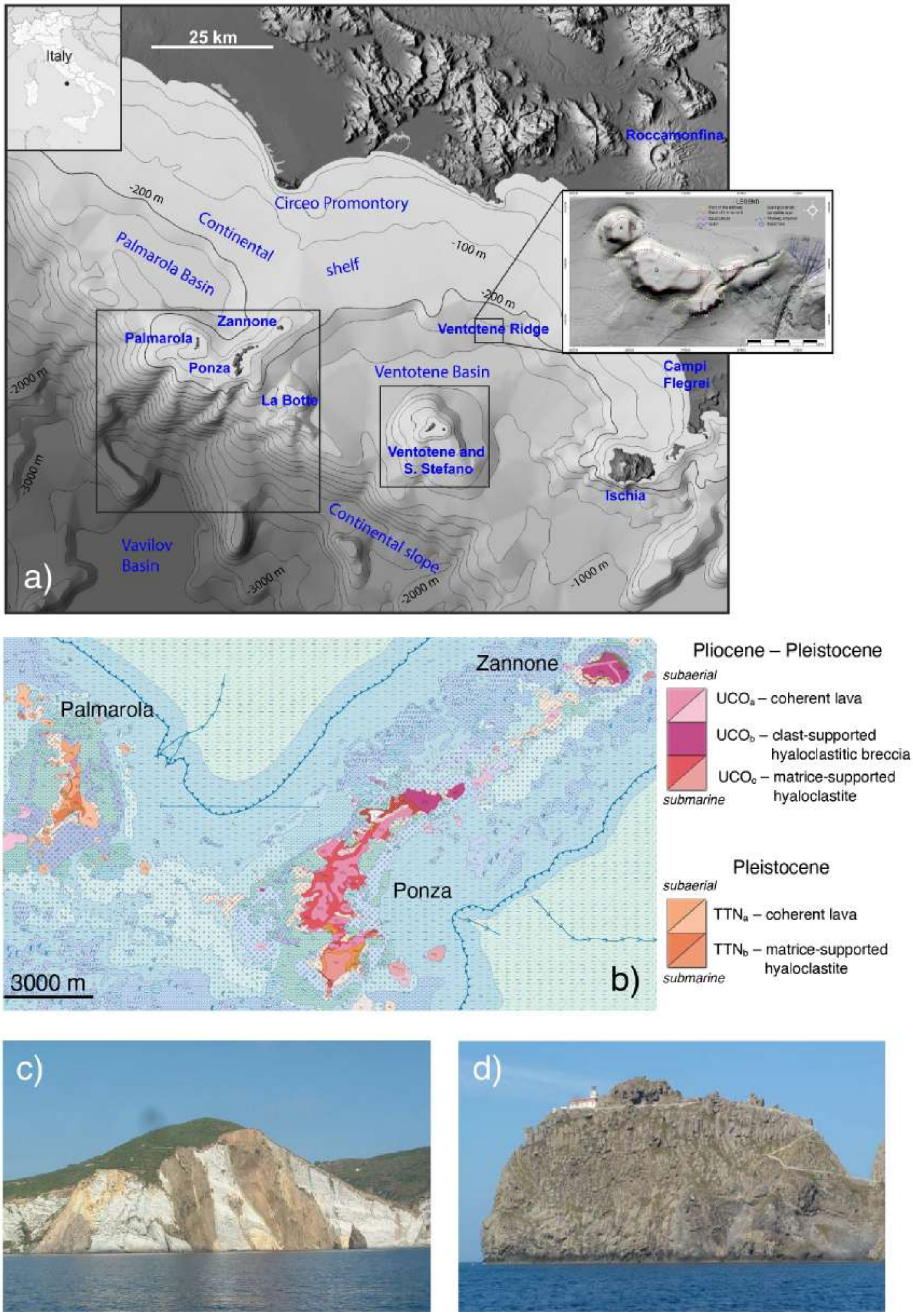


Fig. 16 - (a) Location of the Pontine Archipelago in the central Tyrrhenian Sea. The inset indicates the morphological high of Ventotene Volcanic Ridge (modified after Conte et al., 2016 and 2020); (b) Selected portion of the 1:50,000 geological map of Foglio 413-Borgo Grappa (reported at reduced scale). In the simplified map legend are reported only the main volcanic units; (c) Pliocene: hyaloclastite cryptodoma intruded by dykes at Ponza; (d) Pleistocene: trachytic lava dyke at Punta della Guardia, Ponza, showing cooling structures (vertical and curved columnar jointing).

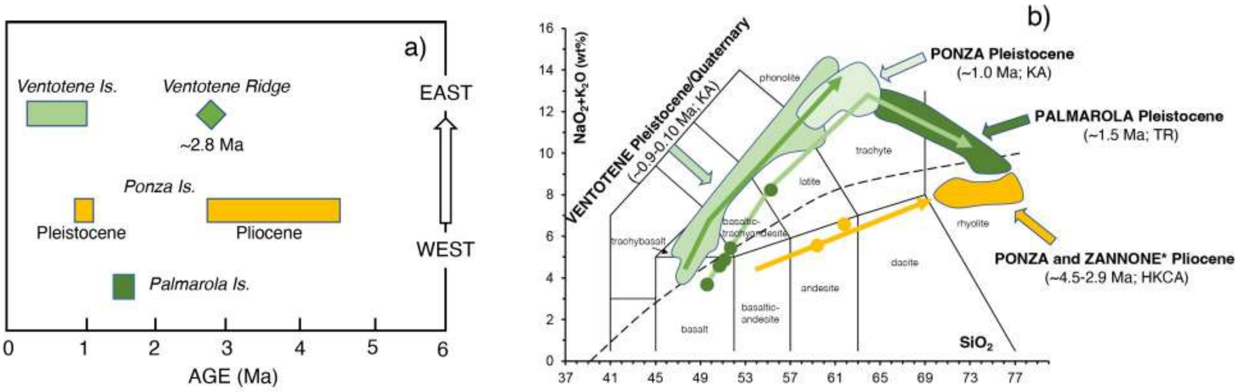


Fig. 17 - (a) Chronology and serial affinity of Pliocene-Quaternary volcanism of the Pontine Archipelago and Ventotene Volcanic Ridge. Yellow, HKCA; pale green, KA; dark green, TR. See the text for the abbreviation; (b) Total Alkali-Silica (TAS; Le Maitre et al., 2002) classification diagram reporting the whole range of compositions of volcanic rocks from Pontine Archipelago. The arrows indicate the liquid lines of descent simulating fractional crystallization trends leading to the felsic rocks constituting the islands from the relatively undifferentiated parental melts recovered in the islands and surroundings. Specifically, a trachybasalt similar to that of Ventotene led to the KA trachyte and trachyte-phonolite of Ponza and Ventotene, respectively; andesite rocks recovered by dredges in the deep water (yellow fill circles) led to the HKCA Pliocene rhyolites, and transitional basalts from both deep water (green small fill circles) and Ventotene Volcanic Ridge led the TR Pleistocene peralkaline rhyolites from Palmarola (Conte et al. 2016, 2020).
* The age of Zannone Island is deduced by the morphological continuity with Ponza and the compositional affinities of the products of the two islands. Indeed, since now no absolute dating exists for the Zannone due to the high alteration degree of the volcanic rocks (Dolfi & Conte, 2018).

4.2. The Neapolitan volcanoes: Somma-Vesuvius, Campi Flegrei and Ischia Island

(R.S., R.C., S.d.V., M.A.D.V., R.I., S.M.)

Somma-Vesuvius, Campi Flegrei and Ischia Island are among the most famous active volcanoes in the world and are part of the Neapolitan landscape. The activity of the volcanoes punctuated the human history since Bronze Age, as testified by archaeological findings recognised in the plain and relieves surrounding the volcanic area. The world-famous eruption of 79 CE consigned the Somma-Vesuvius to the history, because of the burying of some important Roman towns like Pompeii and Herculaneum. The Campi Flegrei volcanic field largely contributed to the birth of myths, and was one of the seven gates of Roman hell. Ischia Island was always a favorite place for thermal baths since Roman times.

This brief chapter summarizes the main characteristics of these three world famous volcanoes, resuming their eruptive history, eruptive behavior, composition of their products and finally, the main hazards they pose to the surroundings and distal regions.

4.2.1. Somma-Vesuvius (39 ka-1944 CE)

The Somma-Vesuvius volcanic complex consists of an older volcano dissected by a summit caldera, Monte Somma, and a recent cone, Vesuvius, built within the caldera after the 79 CE “Pompeii” eruption (Cioni et al., 1999). The original Roman name Vesuvius (or Vesbius) was first applied to the old volcano. Starting from the fifth century, chroniclers make mention of Mt. Somma, as the highest (“summa”) peak of the mountain. The new cone grew discontinuously during periods of semi-persistent, low-intensity activity (from Strombolian to violent Strombolian, accompanied by important effusive activity). These periods possibly occurred in the first to third centuries, in the fifth to eighth centuries (after the 472 CE “Pollena” eruption), in the tenth to twelfth centuries, and in 1631-1944 time span (Cioni et al., 2008; Santacroce et al., 2008).

Until relatively recent times, the formation of Somma-Vesuvius (SV) caldera was ascribed to the Pompeii eruption. Roman paintings from Pompeii and Herculaneum prompted Stothers & Rampino (1983) to conclude that, prior to 79 CE, the top of the volcano was asymmetrically shaped, indicating that a Somma-type caldera was already present. The volcanological interpretation of the Roman fresco from Pompeii "Mars and Venus", now at Naples National Archaeological Museum, made by Nazzaro (1997) leaves few doubts about the presence of a pre-existing caldera (Cioni et al., 1999). The SV caldera has a lobate, quasi-elliptical shape with a 5-km-long, east -west major axis (Fig. 18a). The northern rim of the caldera is a well-defined steep wall, with an average elevation of approximately 1000 m. The drainage pattern of the highest portions of the volcano, unaffected by conspicuous deposition of recent products, is clearly radial, suggesting a symmetrical original cone. In this assumption, the shape and size of the ancient Mt. Somma can be constrained, placing the apex of the old volcano around 1900 m elevation and approximately 500 m North of the present Vesuvius crater. Although the general morphology of the volcano (Fig. 18a) could be suggestive of the occurrence of lateral collapses of the seaward, western sector (Milia et al., 1998; Ventura et al., 1999; Rolandi et al., 2004), field evidence discards such a possibility at least for the last 20,000 years (Sulpizio et al., 2008).

About 22 cal ky BP, the mainly effusive activity of Mt. Somma changed into largely explosive (Fig. 18b). At least four high-magnitude Plinian eruptions occurred, staggered with minor events covering a large range of magnitude and intensity. High-intensity, explosive eruptions sporadically occurred, the two largest being the Subplinian events of 472 CE (also known as Pollena eruption; Rosi & Santacroce, 1983; Sulpizio et al., 2005) and 1631 CE (Rosi et al., 1993; Bertagnini et al., 2006).

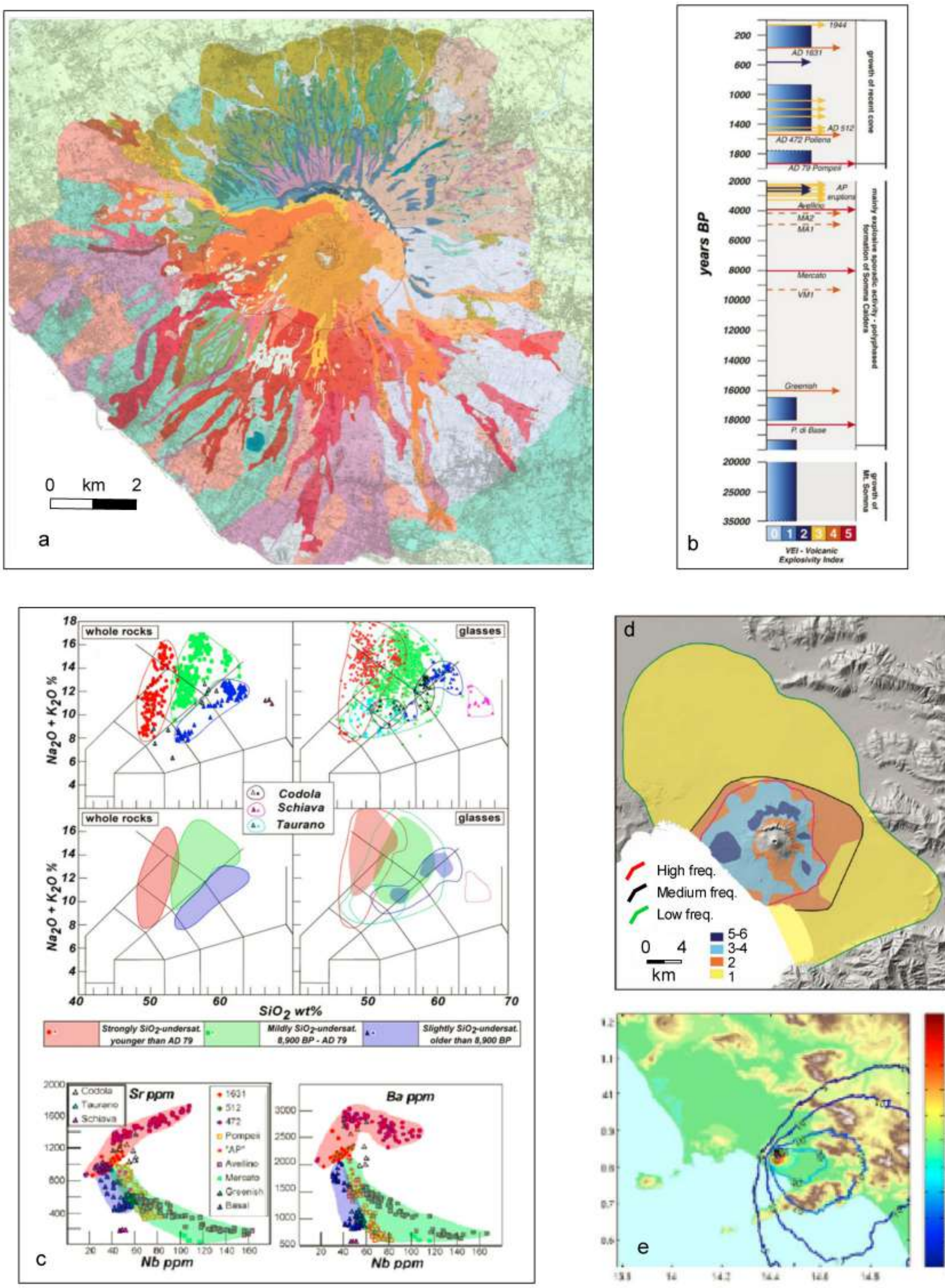


Fig. 18 - (a) Geological map of Somma-Vesuvius volcano (Santacroce et al., 2003; Sbrana et al., 2020). For the legend the reader has to refer to the original publication; (b) Chronogram of Somma-Vesuvius activity. Arrows refer to explosive eruptions, length and colour reflect the estimated VEI. Blue boxes show recorded or inferred periods of persistent mild Strombolian and effusive activity, punctuated by VEI 2-3 explosive eruptions; (c) Variation of Somma-Vesuvius rocks and glasses from the major explosive eruptions within Total Alkalis vs. Silica (TAS) diagram (Santacroce et al., 2008). The variations of Nb, Sr and Ba of major explosive eruptions of SV discriminate the three groups of rocks, and are peculiar for most of single eruptions. Note the Pompeii trend in the Sr vs. Ba plot, completely out of the “mildly silica-undersaturated” field (from Santacroce et al., 2008); (d) PDC inundation frequency for the main eruptions of Somma-Vesuvius during the last 22 ka. The map shows areas that relate to high, medium, and low frequency of PDC inundation during the last 22 ka of activity (from Gurioli et al., 2010); (e) Probability hazard map for fallout load exceeding 300 kg/m^2 for large explosive eruptions at Somma-Vesuvius (from Sandri et al., 2016).

The most recent period (1631-1944) was characterised by summit or lateral lava effusions and semi-persistent, mild explosive activity (small lava fountains, gases and vapour emission from the crater) interrupted by pauses lasting from months to a maximum of seven years (Santacroce, 1987; Arrighi et al., 2001; Fig. 18b). Lateral activity was minor and, starting from the 79 CE, the few eruptive fissures were confined to the western and southern sectors of the volcano (Cortini & Scandone, 1982; Santacroce, 1987; Principe et al., 2004; Acocella et al., 2006).

SV activity has been characterised in the last 20 ka by the eruption of Potassic and Ultra-Potassic magmas spanning a wide compositional spectrum and showing an increase with time of both K_2O content and silica undersaturation (Fig. 18c). The shift from Potassic magmas (represented by the trachytes and latites erupted in the period from the Pomici di Base to the Greenish Pumice) to K-rich compositions occurred before 8 ka. The products younger than 8 ka have been characterised by increasing alkalinity, and the products of the last 2 ka of activity (following the 79 CE Pompeii Pumice) show the most alkali-rich compositions and the lowest SiO_2 content of the whole set of the erupted products.

Slightly silica - Undersaturated (to saturated) magmas (blue field; Fig. 18c)

Two large eruptions related to Somma-Vesuvius activity occurred in the period preceding the Mercato Eruption: the Subplinian, 19,000 cal yr BP “Greenish Pumice” and the Plinian, 22,000 cal yr BP “Pomici di Base”. The most relevant common geochemical signature of the rocks of all these eruptions, related to their K-trachytic highly evolved composition, concerns the higher silica and lower alkali contents (with respect to rocks with comparable degree of evolution of the other two groups), as well as their moderate Sr and Ba contents (Fig. 18c). As a whole, the geochemical features of these rocks are coherent with evolutionary trends of K-basaltic liquids initially driven by crystallization of mafic phases and plagioclase and later involving K-feldspar fractionation.

Mildly silica - Undersaturated magmas (green field; Fig. 18c)

This group consists of the pyroclastic products of three Plinian eruptions (79 CE “Pompeii”, 3910 cal yr BP “Avellino” and 8,900 cal yr BP “Mercato”) and at least six other explosive eruptions occurred between the Avellino and Pompeii events (AP1 to AP6; Andronico & Cioni, 2002). Differently from the Mercato deposits, characterized by a strong compositional homogeneity all along the whole eruptive sequence, the products of the other eruptions of this group present important compositional variations, from the most evolved products at the base to the least evolved toward the top. The Mercato and Avellino eruptions (first-erupted white pumice) are characterized by the emission of the most evolved products of Somma-Vesuvius (CaO<2.0%, Nb>100 ppm; Zr>700 ppm, Th up to 100 ppm).

Strongly silica - Undersaturated magmas (pink field; Fig. 18c)

This group comprises all the products erupted after the 79 CE Pompeii eruption. The most relevant common geochemical signatures of these rocks concern their lower silica and higher alkali contents with respect to rocks with comparable evolution from the other two groups, as well as their very high Sr and Ba contents (Fig. 18c). These features are consistent with evolutionary trends (mostly fractionation within a periodically supplied magma chamber) of K-tephritic liquids dominated by crystallization of leucite and mafic minerals, and characterized by a minor role of plagioclase and the absence of K-feldspar fractionation.

The products of the largest Somma-Vesuvius eruptions are widely dispersed in central Mediterranean area, many of which being important stratigraphic markers for the regional archaeological and paleoclimatic investigations. Codola is dispersed in Tyrrhenian Sea and Adriatic Sea and on land in San Gregorio Magno Basin (Petrosino et al., 2019), Lake Monticchio (Wulf et al., 2004) and Apulia. where acts as a marker for the Palaeolithic successions (Giaccio et al., 2008). Mercato is an important marker for synchronizing the sapropel S1 in Adriatic Sea series with onland Holocene successions of the central-southern Italy and western area of the Balkan Peninsula (e.g., Zanchetta et al., 2011). The Avellino and Pompeii tephra, with their quite opposite dispersal axis, from east to north and south-east, respectively also act as relevant stratigraphic markers for dating and synchronizing marine and and terrestrial successions of the southern-central and southern Italy (Zanchetta et al., 2011; Crocitti et al., 2019; Doronzo et al., 2022).

4.2.2. Campi Flegrei (>80 ka-1538 CE)

Campi Flegrei (CF) is a ca. 14 km wide volcanic caldera located in Southern Italy, in a highly urbanized area including the western part of the city of Naples.

The Campi Flegrei volcanic field (Fig. 19a), includes small volcanic apparatus and several monogenetic tuff rings, tuff cones and rarely cinder cones and lava domes. These volcanoes are exposed outside, on the borders, and within a large polygenetic caldera formed by the eruptions of the Campanian Ignimbrite (CI) and the Neapolitan Yellow Tuff (NYT) (e.g., Armienti et al., 1983; Di Girolamo et al., 1984; Rosi & Sbrana, 1987; Orsi et al., 1992, 1995, 1996, 1999; Cole & Scarpati, 1993; Rosi et al., 1996, 1999; Di Vito et al., 1999; De Vivo et al., 2001; Ort et al., 2003). The northern and western parts of the caldera are above sea level and characterized by the presence of many dispersed cones and craters, whereas the southern part is principally submarine and extends into the Gulf of Pozzuoli. The oldest rocks date back till ca. 80 ka and are represented by pryroclastic deposits cropping out outside and along the borders of the caldera (Pappalardo et al., 1999; Scarpati et al., 2012). Pre-caldera volcanism mainly generated monogenetic tuff cones and subordinate lava domes widespread also within the city of Naples (e.g., Rosi & Sbrana, 1987; Scarpati et al., 2012). Thick pyroclastic sequences generated from at least 80 ka also occur (Orsi et al., 1996; Pappalardo et al. 1999; (Fig. 19a). The caldera forming large volume CI eruption (Fisher et al., 1993; Rosi et al., 1996) erupted ca. 300 km^3 of trachytic dense rock equivalent magma (DRE). The CI eruption was the largest magnitude eruption to occur in the Mediterranean region during the late Quaternary, and resulted in the formation of a 14-km wide caldera (Rosi & Sbrana, 1987). The eruption began with a Plinian phase, during which a SE-distributed pumice fallout was emplaced. This was followed by a succes-

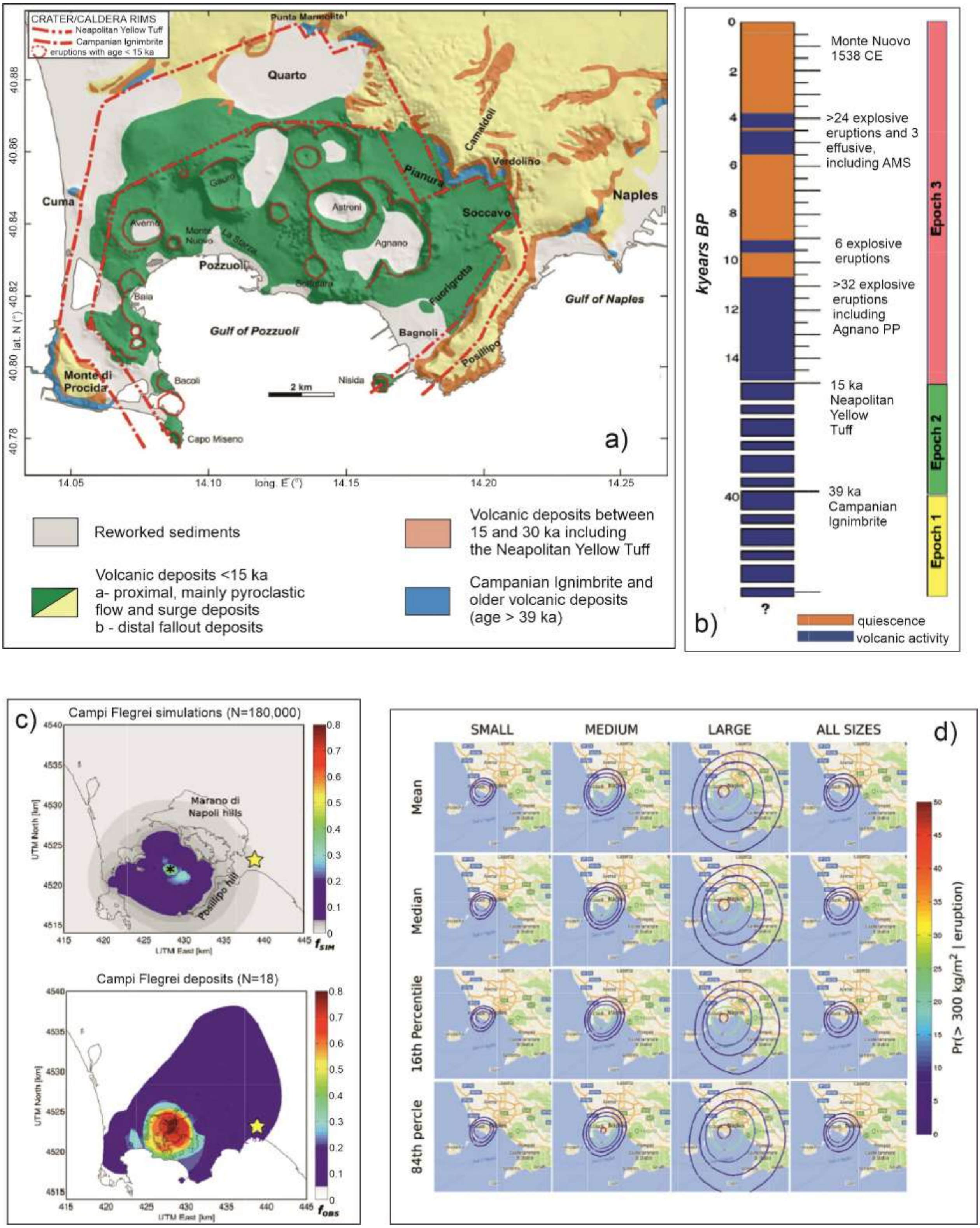


Fig. 19 - (a) Geological sketch map of the Campi Flegrei (from Vitale and Isaia 2014); (b) Chronostratigraphy of Campi Flegrei activity (from Vitale and Isaia, 2014); (c) Frequencies of PDC arrival computed using EC simulations (fSIM) and PDC deposits (fOBS) at Campi Flegrei. Yellow star indicates downtown Naples (from Tierz et al., 2016); (d) Conditional probability maps for a loading threshold of 300 kg/m^2, for size Small, Medium, Large and all sizes (from Selva et al., 2018).

sion of pyroclastic density currents (PDC) that deposited ash and pumice flows and densely-welded ignimbrites that covered the Campanian Plain and surrounding hills (Barberi et al., 1978). Proximal deposits (e.g., Breccia Museo, Piperno) cropping out at the top of the eruption deposits along the caldera margins are interpreted as proximal facies related to the final caldera-forming phase (Rosi & Sbrana, 1987; Rosi et al., 1996). On these deposits, several age data were produced (ca. 37 ka, Deino et al., 2004; ca. 39 ka, De Vivo et al., 2001; ca. 38 ka, Fedele et al., 2008). A recent and high precision ^{14}C and $^{39}Ar/^{40}Ar$ age determination of the CI yielded an age of 39.85±0.14 ka (Giaccio et al., 2017b).

After the CI caldera forming eruption volcanic activity occurred almost exclusively through the formation of pyroclastic deposits from several volcanic centers located within or along the border of the caldera. This period of activity culminated with the second large eruption at CF occurred at 15 ka (Deino et al., 2004) named Neapolitan Yellow Tuff (NYT; Cole & Scarpati, 1993; Wohletz et al., 1995; Fig. 19b), whose volume has been estimated larger than 20 km DRE of latitic to trachytic magma (Orsi et al., 1992, 1995; Scarpati et al., 1993). The eruption was mainly phreatomagmatic producing pyroclastic density currents and subordinate thin fallout layers. The PDC deposits close to CF caldera were largely zeolitized and form the substrate on which lies the city of Naples. Ashes were mainly dispersed toward North-Northeast from the vent area.

The eruptions of the post-NYT period were confined within the structural boundaries of the caldera and comprised at least 70 known events, dominated by low- to medium-magnitude phreatomagmatic-magmatic eruptions with volumes of <0.1 km^3 (Di Renzo et al., 2011; Smith et al., 2011). The Monte Nuovo tuff cone formed as consequence of the most recent eruption in 1538 CE (e.g., D'Oriano et al., 2005; Di Vito et al., 2016). After the NYT eruption, the caldera suffered significant ground deformation phenomena, especially in its central sector, where the uplift is still ongoing.

The primary magmas produced have a mid-ocean ridge basalt (MORB)-like asthenospheric mantle wedge composition, and these are modified by aqueous fluids, oceanic sediments, and continental crust (Tonarini et al., 2004; D'Antonio et al., 2007). The mafic melts at Campi Flegrei are K-basalts (e.g., Webster et al., 2003). These mafic compositions are preserved as melt inclusions in antecrystic Mg-rich olivines and clinopyroxenes in some eruption deposits (e.g., Cannatelli et al., 2007). The composition of the erupted melts range from shoshonitic to phonolitic and trachytic, with the most differentiated compositions dominating (e.g., Mangiacapra et al., 2008; Smith et al., 2011; Tomlinson et al., 2012; Fig. 19c). The phenocrysts in these Campi Flegrei magmas are predominantly plagioclase + K-feldspar + clinopyroxene ± biotite. Magnetite and apatite occur as accessory phases and eruption products occasionally contain olivine or rare feldspathoids.

The SiO_2 and Na_2O contents of the magmas increase with differentiation, whereas CaO, FeO, MgO, and P_2O_5 contents decrease (e.g., Civetta et al., 1991b). There is a noticeable inflection in K_2O melt compositions, denoting K-feldspar-in, at ca. 60 wt% SiO_2 (Fowler et al., 2007; Smith et al., 2011; Tomlinson et al., 2012). The Sr, Ba, and Eu contents behave compatibly, and reflect the significant amount of felspar fractionation. Other REE (excluding Eu), Y, Nb, Zr, Rb, Th, and Ta are all incompatible (e.g., Civetta et al., 1997; Bohrson et al., 2006, Arienzo et al., 2010; Tomlinson et al., 2012). These major and trace element compositions follow an evolutionary trend that could be generated through fractional crystallization of a single parental melt (e.g., Civetta et al., 1991a; D'Antonio et al., 1999; Fourmentraux et al., 2012) but isotopic variations indicate that the melts that erupt are derived from different batches of magma (Pappalardo et al., 1999, 2002; D'Antonio et al., 2007; Di Renzo et al., 2011; Fig. 19c).

Samples from Campi Flegrei eruption deposits suggest that only evolved magmas were erupted in the early history (Pappalardo et al., 1999) and it was only after the last caldera-forming NYT eruption (ca. 15 ka) that more compositionally diverse melts were erupted (e.g., D'Antonio et al., 1999; Smith et al., 2011). The isotope (Nd, Pb and Sr) and occasionally the major and trace element glass compositions of the magmas indicate that the eruptions tap distinct batches of melt, and some have interacted at depth (e.g., Di Renzo et al., 2011). This has been well documented for the large caldera-forming events (e.g., Forni et al., 2018) and for eruptions in the last 15 ka (e.g., Tonarini et al., 2009; Fourmentraux et al., 2012). The general trend between 60 and 10 ka is that Nd and Pb isotopic compositions of the erupted magmas became progressively less radiogenic ($^{143}Nd/^{144}Nd$ - 0.51252 to 0.51236 and $^{206}Pb/^{204}Pb$ - 19.2 to 18.9) while the Sr-isotope composition became more enriched (0.70700 to 0.70864) over time (Pabst et al., 2008; Di Renzo et al., 2011). These changes in the isotopic compositions are consistent with an increase in crustal contamination. However, the Campi Flegrei liquid line of descent and extent of Sr and Pb isotopic heterogeneity is compatible only with very minor assimilation (D'Antonio et al., 2007; Fowler et al., 2007).

The continuous and intense activity of the Campi Flegrei, makes this volcanic system the main source of the late Pleistocene tephra in central Mediterranean area. Y-5, the tephra of the CI, the largest Campi Flegrei eruption, is to great extent the most widespread, and one of the most relevant stratigraphic markers for synchronizing stratigraphic, paleoclimatic and cultural events occured in western Eurasia around 40 ka (e.g., Giaccio et al., 2017b). Though still poorly known in near vent sections, a number of other widely dispersed tephra preceding the CI eruption, including, among other the X-6, X-5 and C-22 marine layers, have been attributed to previously unknown 92-109 ka Campi Flegrei activity (Monaco et al., 2022b). Following the CI eruption, the two other tephrostratigraphic markers, which significantly exceed the regional dispersal, are those of the relatively poorly known Masseria del Monte eruption of ca. 29 ka (Y-3, Albert et al., 2019) and of the Neapolitan Yellow Tuff, mainly dispersed toward North and found until central Europe (e.g., Lane et al., 2011; 2015).

4.2.3. Ischia Island (>150 ka-1302 CE)

The island of Ischia is part of the Phlegraean volcanic district, related to extensional tectonic phases that accompanied the anticlockwise rotation of the Italian peninsula (Ippolito et al., 1973; D'Argenio et al., 1973;

Finetti & Morelli, 1974; Bartole, 1984; Piochi et al., 2005). Ischia is the emerged part of an active volcanic field, which rises more than 1,000 m above the sea floor (Fig. 20a; Orsi et al., 1999; Bruno et al., 2002), along the margin of an E-W trending scarp that borders to the south the Phlegraean volcanic district. The island covers an area of 46.4 km^2 and is morphologically dominated by the Monte Epomeo (787 m a.s.l.) in its central portion, and by the NE-SW Monte Vezzi - Monte Cotto alignment in the SE corner.

Volcanism at Ischia dates back to more than 150 ka and continued, with centuries to millennia of quiescence, until the most recent eruption occurred in 1302 CE (Vezzoli, 1988; Sbrana & Toccaceli, 2011; Sbrana et al., 2018; Fig. 20b). The oldest exposed rocks, aged between 150 and 75 ka, are lava flows and/or lava domes, and a sequence of pyroclastic deposits with intercalated paleosols, mainly exposed in the southern part of the island (Scarrupata di Barano Formation in Vezzoli, 1988; Ancient Ischia Synthem in Sbrana & Toccaceli, 2011; Phase 1 in Sbrana et al., 2018; l.c. volcanics older than 75 ka of Figs. 20b and c). Almost in the same time span, a prevailing effusive activity determined the emplacement of a series of small trachytic and phonolitic lava domes, which are presently exposed at the periphery of the island (Fig. 20a).

A very intense period of explosive activity followed between 74 and 55 ka. During this period, many volcanic vents were active mainly in the southern sector of the island, likely producing the highest magnitude eruptions recorded at Ischia. The deposits of at least 10 explosive eruptions, fed by phonolitic to trachytic magmas, generated pyroclastic density currents, fallout deposits, block and ash flows from collapsing lava domes, explosion breccias and hydromagmatic dilute and turbulent pyroclastic density currents (Brown et al., 2008; Rifugio di San Nicola Synthem in Sbrana & Toccaceli, 2011; Phase 2 in Sbrana et al., 2018). The activity of this period culminated with the caldera forming eruptions of the Monte Epomeo Green Tuff (MEGT) between 60 and 50 ka (Brown et al., 2008; Sbrana & Toccaceli, 2011; Sbrana et al., 2018; Fig. 20b).

This tuff consists mostly of trachytic ignimbrites that partially filled the caldera depression in a submarine environment, and were also emplaced on land outside its margins. The caldera depression was later the site of marine sedimentation, which formed a succession of clays, tuffites, sandstones and siltstones by the reworking of MEGT and sedimentary supply from the mainland (Sbrana & Toccaceli, 2011). After the MEGT eruptions, volcanism continued with a series of hydromagmatic and magmatic explosive eruptions up to 33 ka. Most of the vents were located along the present SW and NW offshore of the island. Volcanism in this period was fed by latitic and shoshonitic magmas (Brown et al., 2014; Phase 3 in Sbrana et al., 2018), whose injection into the system triggered the asymmetric resurgence of the caldera floor, generating the Monte Epomeo block (Orsi et al., 1991). After an about 5 ka long period of quiescence, volcanism resumed at ca. 28 ka. The beginning of this period of activity was marked by the eruption of shoshonitic magmas, isotopically distinct from those characterizing the previous eruptions, suggesting the arrival of a new batch of magma into the feeding system of Ischia, followed by its differentiation and mixing with the resident magma (Poli et al., 1987; Civetta et al., 1991b; Casalini et al., 2017). Volcanism continued sporadically until 18 ka (or 13 ka, Phase 4 in Sbrana et al., 2018), with the emission of shoshonitic to alkali-trachytic magmas that fed effusive and both hydromagmatic and magmatic explosive eruptions, with the emplacement of lava flows and Strombolian fallout deposits, and the construction of small tuff-cones and scoria-cones, exposed along the SE and SW coasts of the island (Fig. 20c; Campotese Subsynthem in Sbrana & Toccaceli, 2011).

The radiometric dating of the outcropping volcanic products suggests that between 18 and 10 ka there was a period of almost complete quiescence (Vezzoli, 1988), whereas other interpretations (Sbrana & Toccaceli, 2011; Sbrana et al., 2018) seem to indicate the occurrence of some eruptions in this timespan. In any case, all the authors agree in considering the interval between 18 and 10 ka as a period of quiescence or very reduced volcanic activity (Fig. 20b). A period of more sustained activity at Ischia started at ca. 10 ka, and was fed by mainly trachytic and subordinately latitic magmas. These are characterized by a Sr isotopic ratio lower than those of the previous periods, leading to hypothesize the arrival of a new, geochemically distinct magma into the system. Volcanism was mainly concentrated around 5.5 ka and after 2.9 ka, with almost all the vents located in the eastern part of the island (Fig. 20c). Only a few vents are located outside this area, along regional fault systems or along the margins of the resurgent block: these vents generated a multi-vent lava field in the NW corner of the island, a pyroclastic sequence, exposed to the SW, and a lava dome in the N sector of the island. Since 2.9 ka, about 34 effusive and explosive eruptions took place: the effusive eruptions emplaced lava domes and high-aspect ratio lava flows, while the explosive eruptions, both magmatic and phreatomagmatic, generated tuff cones, tuff rings and variably dispersed pyroclastic-fall and-current deposits, which had a very variable impact on both the island's environment and human settlements (de Vita et al., 2010 and references therein; Sbrana & Toccaceli, 2011; de Vita et al., 2013; Phases 5 and 6 in Sbrana et al., 2018). As volcanism was not continuous since 10 ka, but characterized by the alternation of centuries of quiescence and periods of very intense activity, it has been hypothesized that repeated episodes of magma intrusions occurred intermittently (Tibaldi & Vezzoli, 2004; de Vita et al., 2006; Vezzoli et al., 2009; de Vita et al., 2010), likely accompanied by resurgence. Even if, since the last eruption, no evidence of renewal of uplift due to fault reactivation has been recorded in concurrence with more recent minor mass movements, the magmatic system of Ischia has to be considered still active, as testified by intense volcanism in historical times and widespread fumaroles and thermal springs.

The Ischia volcanics belongs to the low-K series of the Roman comagmatic province, ranging in composition from shoshonite to latite, trachyte and phonolite; the most abundant exposed rocks are alkali-trachytes (Angiulli et al., 1985; Poli et al., 1987, 1989; Crisci et al., 1989; Civetta et al., 1991; Fig. 20d).

The activity of Ischia volcano is well documented in distal settings (e.g., Wulf et al., 2004; 2008). The most widespread Ischia tephra is the one related to the Monte

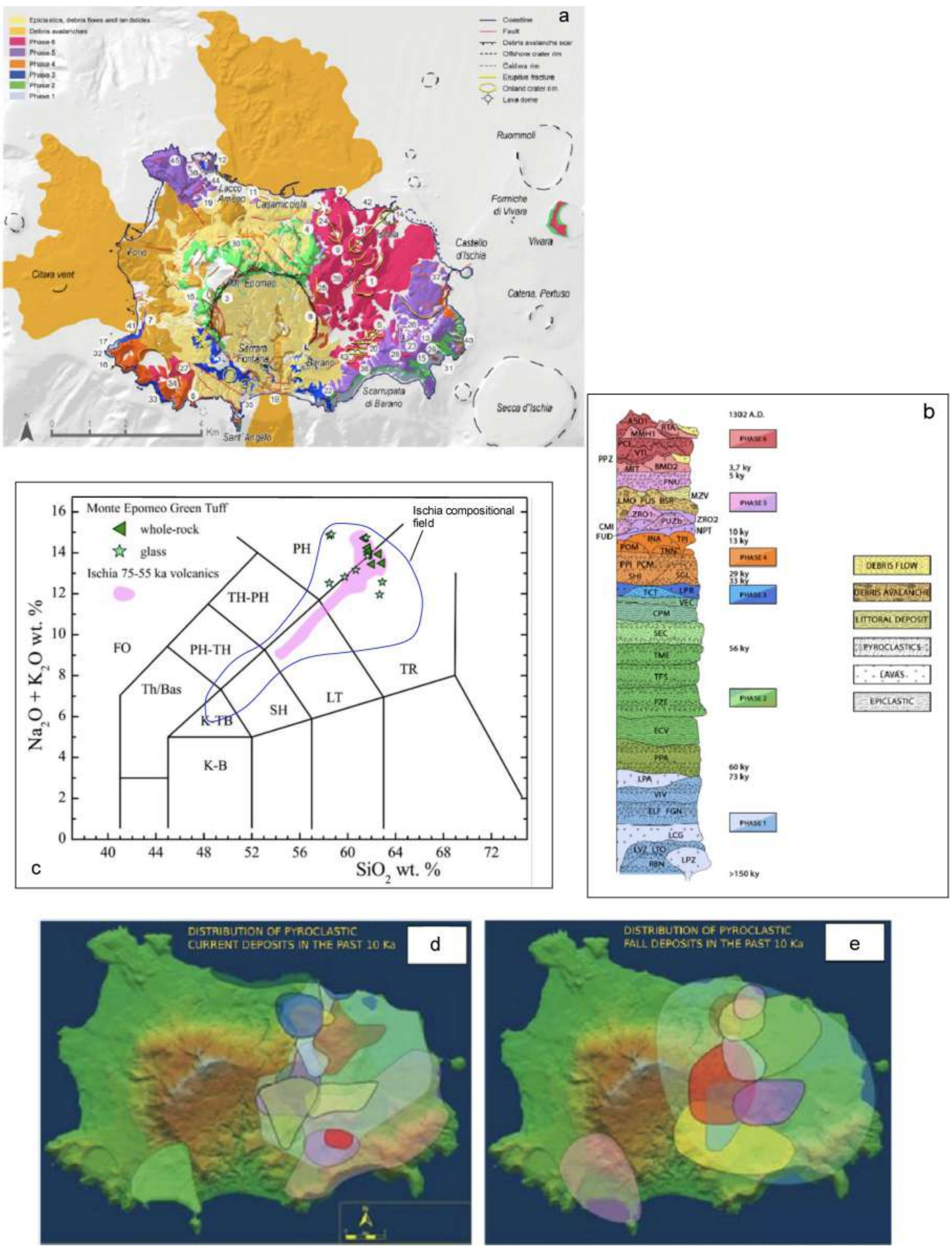


Fig. 20 - (a) Geological and structural sketch map of Ischia volcanic field. Deposits of debris avalanches and volcanic vents are highlighted in marine areas (from Sbrana et al., 2018). For the legend the reader has to refer to the original publication; (b) simplified chronostratigraphic sequence (not to scale) of Ischia volcanic field (from Sbrana et al., 2018); (c) Total alkali vs. silica classification diagram showing the variability of composition for Ischia products (from D'Antonio et al., 2021; Santacroce et al., 2008). K-B = potassic basalt; K-TB = potassic trachybasalt; SH = shoshonite; LT = latite; TR = trachyte (these rock names are relative to a potassic alkaline series); TH/Bas = tephrite/basanite; PH-TH = phonotephrite; TH-PH = tephriphonolite; PH = phonolite; (d) distribution of PDC in the last 10 ka (from Selva et al., 2019); (e) distribution of pyroclastic fall-out deposits for all the eruptions in the last 10 ka (from Selva et al., 2019).

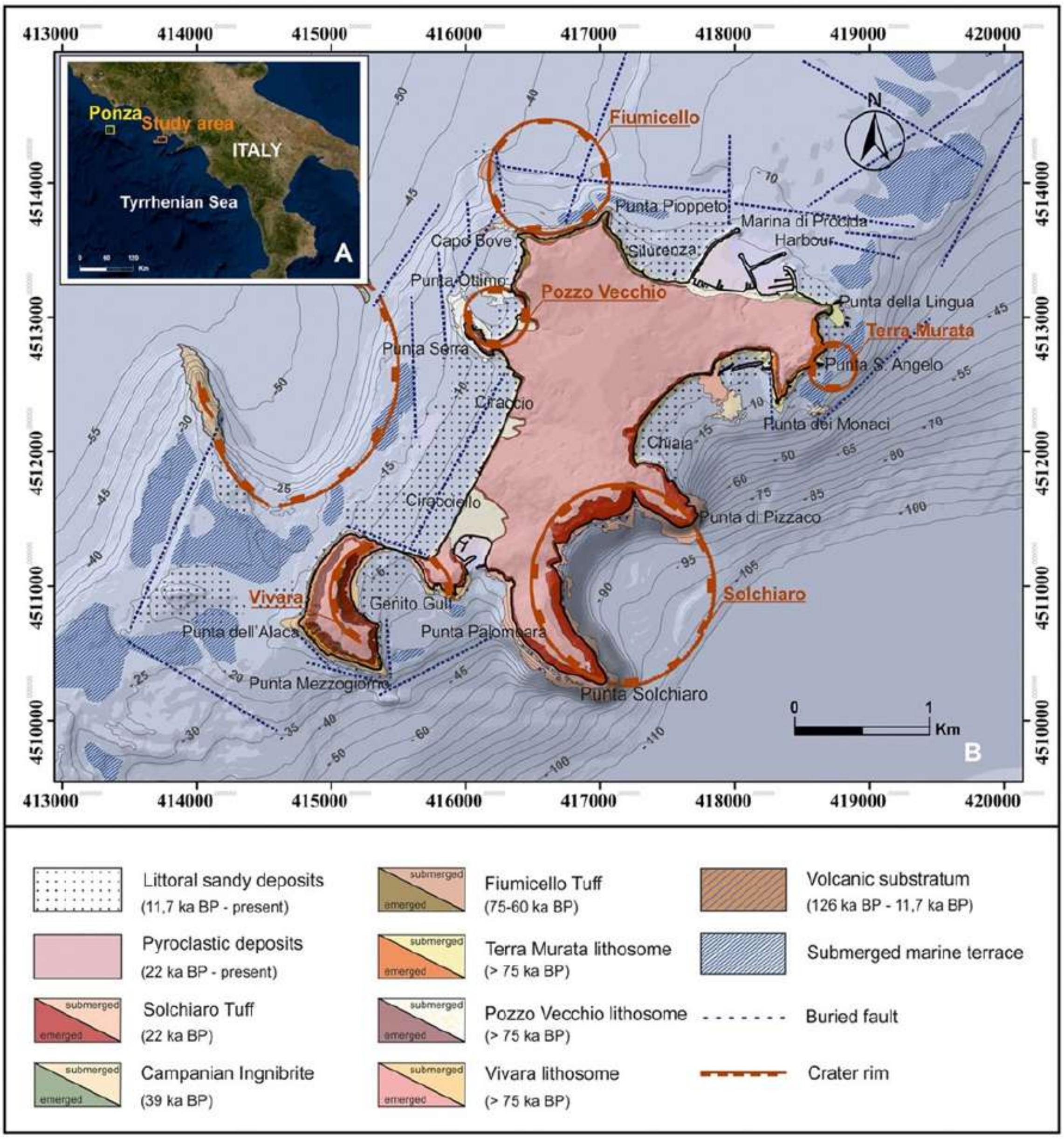


Fig. 21 - Geological map of Procida Island with the chronological classification of the main volcanic events and the location of the main coastal sites (from Aucelli et al., 2022, modified after Fedele et al., 2012). The position of the submerged marine terraces was obtained from Putignano et al. (2014).

Epomeo Green Tuff eruption of ca. 56 ka (Tomlinson et al., 2014), matching the layer TM-19 of Monticchio (Wulf et al., 2004), S16 of San Gregorio Magno (Petrosino et al., 2019) and TF-7 of Fucino (Giaccio et al., 2017a), and a still unknown, slightly older, ca. 60 ka, eruption that matches the widespread marine layer Y-7 (Keller et al., 1978) and the Monticchio tephra TM-20 (D'Antonio et al., 2021). Several tephra layers that are compositionally attributable to Ischia, preceding both MEGT and the unknown Y-7-related eruption, found, e.g., in the Tyrrhenian Sea (Paterne et al., 2008) and Fucino (Monaco et al., 2022a), extend the activity of this volcano at least to 200 ka.

4.2.4. Procida Island

The volcanic island of Procida formed due to several monogenetic explosive eruptions that led to the formation of five volcanic centres characterizing the emerged-submerged morphology of the island: Vivara, Pozzo Vecchio, Terra Murata, Fiumicello, and Solchiaro (Fig. 21; Aucelli et al., 2022). These tuff rings, except the tuff cone of Terra Murata, are SW-NE oriented following a fault line crossing the centre of Procida and the nearby island of Ischia (e.g., Pescatore & Rolandi, 1981).

De Astis et al. (2004) tentatively constrained the beginning of the volcanic activity of Procida Island around 70 ka and several authors (Alessio et al., 1976; Lirer et al., 1991; Fedele et al., 2012) identified the last volcanic episode of the area with the Solchiaro eruption, occurred 22 ka (Fedele et al., 2012). The different volcanic episodes, which led to several volcano-tectonic collapses, modified the original structure of the island together with the activity of the main fault systems (Putignano & Schiattarella, 2010). The deposits deriving from these activities, mainly made up of stratified to massive ash levels, are covered by a tephra succession, cropping out along the SW flank of Vivara and Pozzo Vecchio and partially at Terra Murata, and interbedded with the volcanic formation of Pignatiello (55-74 ka; Fedele et al., 2012).

Procida rocks range in composition from basalt to shoshonite and trachyte. The presence of a compositional gap in the range SiO2=54-59 wt % is evidence of magma bimodality, suggesting that the feeding magmatic system was formed by at least two different reservoirs located at different depths (De Astis et al., 2004).

4.2.5. Hazard assessment

The Neapolitan area represents one of the highest volcanic risk areas in the world due to the presence of three active volcanoes. In the last decades, many studies assessed tephra fall hazard from them combining field data and numerical simulations. Initially, conditional ash load probability maps for one or few reference volcanic scenarios were provided (e.g., Barberi et al., 1995; Cioni et al., 2003; Macedonio et al., 2008; Costa et al., 2009; Folch & Sulpizio, 2010; Sulpizio et al., 2012). The choice of selecting a single scenario was very useful to support civil protection emergency plans (e.g., Emergency Plan of Vesuvius; DPC 2015) and it is computationally less expensive and more feasible for near-real-time applications (e.g., Selva et al., 2014). However, this strategy was not appropriate to achieve an unbiased Probabilistic Volcanic Hazard Assessment (PVHA; Marzocchi et al., 2010; Selva et al., 2010; Sandri et al., 2016) since the uncertainty only accounted for wind conditions (e.g., Macedonio et al., 2008; 2016).

To overcome this limitation, the analysis of the intra-scenario variability is certainly more complete, allowing a reduction of the epistemic uncertainty (e.g., Selva et al., 2014; Sandri et al., 2016; Titos et al., 2022) in both long- and short-term hazard assessments (e.g., Selva et al., 2010; Sandri et al., 2012; 2014; Selva et al., 2014; Thompson et al., 2015; Constantinescu et al., 2016).

In more recent PVHAs at the Neapolitan volcanoes, new sampling procedures have been set to fully explore the intra-size-class aleatory variability comparing the results with the classical approach based on reference volcanic scenarios, in the case of proximal/medial areas and large tephra loads from Somma-Vesuvius and Campi Flegrei (Sandri et al., 2016) or by using ensemble modeling of submarine eruptive vents and tephra total grain-size distributions considering variable mass fraction and aggregates for ash (Selva et al., 2018). Moreover, Montesinos et al. (2022) have developed a new Bayesian Event Tree based on high performance computing applied to Campi Flegrei, aiming to robust and unbiased short- and long-term PVHA on a large-scale (thousands km) and high-resolution (about 2 km) domain to be used by Civil Protection agencies, aviation companies and other stakeholders.

A novel contribution in the quantification of volcanic hazard regards the estimation of the present state of the Neapolitan volcanoes based on a simple physics-based statistical model that satisfactorily fits the eruptive history of each volcano (Selva et al., 2022). The model is compatible with existing data (including isolated events and long repose periods) and accounts for two activity regimes (high-low) able to describe the temporal modulations in eruptive activity and to provide a homogeneous quantification of the probability of eruption, considering the state of the volcano and the possible transitions.

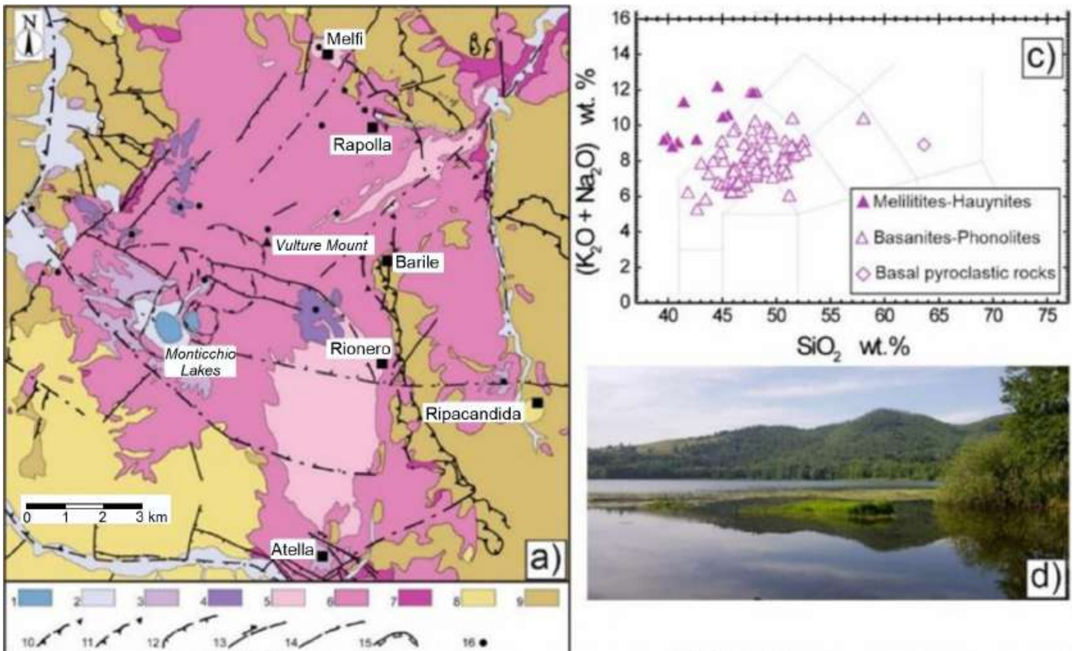
More recently, Massaro et al. (2023) proposed a new methodology for a probabilistic long-term tephra fallout hazard assessment in Southern Italy from the active Neapolitan volcanoes (Somma-Vesuvius, Campi Flegrei and Ischia). By means of thousands of numerical simulations they quantify the mean annual frequency with which the tephra load at the ground exceeds critical thresholds in 50 years. The output hazard maps account for changes in eruptive regimes of each volcano and are also comparable with those of other natural disasters in which more sources are integrated.

For pyroclastic density currents, Gurioli et al. (2010) provided the high, medium and low inundation frequency from the main eruptions of Somma-Vesuvius during the last 22 ka of activity (Fig. 18d). More recently, quantitative PVHAs were carried out by Tierz et al. (2016) for Somma-Vesuvius and Campi Flegrei testing the energy cone model (Fig. 19d) and by Selva et al. (2019) for Ischia Island (Fig. 20e-f).

4.3. The Mt. Vulture and Monticchio Lakes nested volcanoes (>730-130 ka) (S.C., P.P., J.P.S.-B.)

The Monte Vulture and Monticchio nested volcanoes with few small eccentric explosive monogenetic centres along the Ofanto valley and at Ripacandida make up the Lucanian Magmatic Province (Fig. 1). The volcanism of this magmatic province is located East of the Apennine system, at the outermost edge of the fold and trust belt. The volcano lies on a structural high made up of Meso-Cenozoic units of the substratum overlain by Pliocene continental sediments (Schiattarella & Beneduce, 2006) and is situated on a deep transfer fault (Trinitapoli-Paestum Line in Ciaranfi et al., 1983; Vulture Line in Schiattarella et al., 2005).

The volcanic activity at Vulture-Monticchio nested volcanoes (Fig. 22a) developed entirely within the Pleistocene starting at about 0.74 Ma (Villa & Buettner, 2009). A reconstruction of the eruptive history of the volcano, after the pioneering work of Hieke Merlin (1967), was carried out by Principe & La Volpe (1989a), who recognized several markers useful to build a stratigraphic framework. The eruptive history has been systematically reinvestigated in the frame of the CARG project by Giannandrea et al. (2006), who adopted USBU systematics to map the primary volcanic products and the sedimentary deposits of the several fluvial/lacustrine successions that characterise the perivolcanic area (Fig. 22b). They highlight the strong tectonic control on the whole of the Vulture activity and identify erosional phases, epiclastic sediments and/or paleosoils which testify to the occurrence of variably long quiescence phases. The very first dated products pertain to the Fara d'Olivo sub-synthem (Giannandrea et al., 2006) and are represented by two main ignimbrite units (Ignimbrite A and B of Crisci et al.,



		SYNTEM	SUBSYNTEM			
SUPERSYNTEM	Monticchio	Monticchio Lakes	Casa Rossa	Lago Piccolo Lago Grande Piano Comune	Serra di Braida	
		Valle dei Grigi - Fosso del Corbo	Case Lopes	Masseria di Cuscito	Imbandina	
	Vulture	Melfi	Castello di Melfi	Piano di Croce	Gaudo	Solagne Arcidiaconata
		Barile	Ventaruolo	Vulture – S. Michele	Rionero	Toppo San Paolo
		Foggianello	Fara d'Olivo	Campanile	Spinoritola	b)

Fig. 22 - (a) Detailed geological sketch map of the Vulture volcanic products. Legend: 1) Monticchio Lakes; 2) Landslides, alluvial and lacustrine deposits; 3) Laghi di Monticchio Synthem (Upper Pleistocene); 4) Valle dei Grigi - Fosso del Corbo Synthem (Middle Pleistocene); 5) Melfi Synthem (Middle Pleistocene); 6) Barile Synthem (Middle Pleistocene); 7) Foggianello Synthem (Middle Pleistocene); 8) Fiumara di Atella Supersynthem (Upper Pliocene-Lower Pleistocene?); 9) Cretaceous to Miocene Apennines units; 10) thrusts; 11) reverse faults; 12) normal faults; 13) strike-slip faults; 14) high-angle faults; 15) crateric rim; 16) secondary volcano eruptive centres. (b) Unconformity bounded stratigraphic units forming the whole volcanic products of the Vulture volcano. Modified from Schiattarella et al. (2005) and Giano et al. (2022); (c) TAS classificative diagram (Le Maitre et al., 2002) for the Vulture volcanic rocks (modified from Conticelli et al., 2010); (d) View of the Lago Grande di Monticchio (available at fondoambiente.it/luoghi/monticchio?ldc).

1983), for the lowermost of which an $^{40}Ar/^{39}Ar$ age of 739±12 ka has been determined (Villa & Buettner, 2009). These ignimbrites, to which a caldera collapse has been ascribed (Principe & Giannandrea, 2006), unconformably overlay the deposits of an older explosive activity (Campanile subsynthem, Giannandrea et al., 2006). A subaerial erosional phase, with the associated epiclastic sediments, marks the passage to the Barile synthem, to which the Toppo San Paolo subsyntem, a lava dome, on the NW rim of the Vulture, belongs, together with the Rionero subsyntem, a complex sequence of pyroclastic fall and pyroclastic flow deposits. Among these deposits, the fallout of Masseria Boccaglie (De Fino et al., 1982) has the prominent role of a stratigraphic marker (marker M3 of La Volpe & Principe, 1989) younger than 720±25 ka (Buettner et al., 2006). Starting from the Rionero, but mainly in the Vulture-San Michele subsynthems (Principe & Giannandrea, 2006) the main stratovolcano was built up between ca. 670 and 600 ka (Brocchini et al., 1994; Villa & Buettner, 2009) by alternating mildly explosive and effusive activity, respectively emplacing block and ash flow with minor pyroclastic fall deposits, and thin lava flows. A phreatomagmatic eruptive phase (Ventaruolo subsynthem) marks the beginning of an intense phase of destruction of the stratovolcano, occurred mainly during the Melfi synthem deposition, in which parasitic vents along the north and north-eastern side of the volcano produced lava flows with the emission of the Melfi haüynophire (557±7 ka, Bonadonna et al., 1998; 573±4 ka, Villa & Buettner, 2009) and Piano di Croce lavas. The volcanic and tectonic activity of the Vulture area has been sealed by a thick paleosoil (Schiattarella et al., 2005), which La Volpe & Principe (1989) had identified as marker M18. The last deposits, grouped in the Monticchio supersynthem by Giannandrea et al. (2006), were erupted starting from ca. 500 ka (Buettner et al., 2006 and references therein) from a series of monogenetic centers, and mostly show evidence of phreatomagmatic activity. This supersynthem also includes the ash tuff deposits of the two maar-like centres of Lago Grande and Lago Piccolo di Monticchio (Fig. 22d). The recommended age of the Lago Piccolo subsynthem, which gives the youngest $^{40}Ar/^{39}Ar$ age determined so far for the Vulture products, is 141±11 ka (Villa & Buettner, 2009).

The bulk of the erupted products mostly belong to a basanite-tephrite-phonolite series (Fig. 22c), with mafic products (basanites and tephrites) dominating in both lavas and pyroclastic rocks (e.g., De Fino et al., 1986; Beccaluva et al., 2002). More silica undersaturated rocks, though minor in volume, are typical of this volcanic complex (Hieke Merlin, 1964, 1967). Melilitites, haüynites, haüyne-bearing leucitites are also found at the end of the Monte Vulture volcanic activity in the form of dykes, plugs and lava flows in the north-western flank of the volcano. On the other hand, trachyphonolites are found as dykes in the neighboring of the volcanic complex, and may be representatives of an earlier, and less silica undersaturated stage of activity (Melluso et al., 1996; Beccaluva et al., 2002).

A small carbonatitic lava flow has been also found at Toppo del Lupo (e.g., Stoppa & Principe, 1997; D'Orazio et al., 2008; Stoppa et al., 2008). Within the hydromagmatic units of the Monticchio supersynthem, round lapilli tuff units with ejecta of intrusive carbonatites and mantle nodules are visible (e.g., Jones et al., 2000; Rosatelli et al., 2000; Downes et al., 2002). K_2O contents of the volcanic products decrease with time passing from Monte Vulture to Monticchio volcanoes. Sodic compositions are also found among lapilli tuffs of the Monticchio lake activity (Avanzinelli et al., 2008, 2009).

Two distal tephra attributed to Vulture activity and tentatively correlated with ignimbrites A and B of Crisci et al. (1983) despite the incomplete age overlap, were found embedded in the Montalbano Jonico marine sequence (Petrosino et al., 2015).

4.4. Aeolian Islands (1.3 Ma-Present)

(F.L., G.D.A., R.D.R., P.D., F.F., L.F., E.N., M.P., R.S., C.R., C.A. T., M.V.)

4.4.1. General background

The Aeolian Islands arc is the most active volcanic structure in the Mediterranean area, and is composed of seven volcanic islands (Alicudi, Filicudi, Salina, Lipari, Vulcano, Panarea, Stromboli) and several seamounts rising ca. 2000-3000 m above the seafloor in the Southern Tyrrhenian Sea (Romagnoli, 2013; Romagnoli et al., 2013) (Fig. 23a). They are arranged in an articulated, tectonically-controlled arc-shaped structure, crossed in its central sector by the NNW-SSE lithospheric Tindari-Letojanni fault system, which has developed in a complex subduction-type scenario during the Quaternary (Ventura, 2013). The oldest products from the submarine sectors are dated to 1.3 Ma (Beccaluva et al., 1985), whereas the age of emergent islands ranges between 270-250 ka (Leocat, 2011) and the Present (Fig. 23b). Different stages of volcanic activity through time are established on the basis of the available radiometric ages and stratigraphic relationships with marine terraces and tephra layers of Campanian and/or Aeolian origin (Lucchi et al., 2013a and references therein). The Aeolian Islands volcanoes are typically characterized by a large variety of volcanic landforms and rock types which mostly depends upon the wide compositional range observed in the erupted products from basalts to rhyolites with calcalkaline (CA) to potassic signature (De Astis et al., 2013; Forni et al., 2013; Francalanci et al., 2013; Lucchi et al., 2013a, b, c, d, e; Nicotra et al., 2020) (Fig. 23c), and the corresponding eruption types (from Strombolian/Hawaiian to Subplinian), under the direct control of regional fault systems and recurrent calderas and lateral collapses. The western Aeolian volcanoes (Salina, Filicudi and Alicudi) are considered extinct, and currently active to quiescent volcanoes (Vulcano, Stromboli, Panarea, Lipari) are basically restricted to the eastern and the outer portions of the sector of the Aeolian Islands arc. Major explosive eruptions involving more evolved magmas occurred during the last 75 ka on Stromboli (Petrazza Tuffs), Salina (Grey Porri Tuffs, Lower and Upper Pollara tuffs) and Lipari (Monte Guardia, Vallone del Gabellotto, Monte Pilato) producing widespread tephra layers recorded in the Tyrrhenian, Ionian and Adriatic deep sea cores and other lake or terrestrial archives in southern Italy (Albert et al., 2017; Meschiari et al., 2020 for references).

Hereafter we describe the most important geological features and chronology of the Alicudi, Filicudi, Salina, Lipari and Panarea volcanic complexes (from West to East), whereas a particular attention will be given to Vulcano and Stromboli in dedicated paragraphs.

The island of Alicudi is the emerged portion of a composite volcano constructed between 106 and 28 ka by lava flows, domes and Strombolian scoriae with a compositional range between CA basalts and high-K CA andesites (Lucchi et al., 2013d). Six Eruptive Epochs were characterized by a mainly central-type volcanism interrupted by periods of dormancy and three summit caldera-type collapses (Lucchi et al., 2013d). The Alicudi rocks are characterized by the most primitive geochemical and isotopic signature of the entire Aeolian archipelago, with a progressive shift toward more acid products due to the evolution of the magma feeding system and polybaric crystal fractionation, and a lower control by extensional regional tectonics (Peccerillo et al., 2013).

The island of Filicudi is the emerged portion of a volcanic complex, result of the superimposition of lava flows, lava domes and coulees and subordinate pyroclastic deposits ranging in composition from CA basalts to (minor) HK CA dacites (Lucchi et al., 2013e). Four Eruptive Epochs developed between 246 and 29 ka, separated by quiescence stages, erosional episodes and partial collapses of portions of the volcanic edifices (Lucchi et al., 2013e). Four partially overlapping stratovolcanoes and scoria cones (Casa Ficarisi, Fossa Felci, Chiumento and Monte Guardia) were mostly controlled by the WNW-ESE regional tectonic trend, with a progressive shifting of activity toward South-East. A variation of compositions toward the dacitic terms was recorded in the lava domes and coulees and explosive eruptions in the south-eastern sector of Filicudi at ca. 64-50 ka (Monte Terrione, Capo Graziano, Monte Montagnola and Case dello Zucco Grande; Lucchi et al., 2013e). Volcanic activity of Filicudi has been regulated by the interaction of different evolutionary processes, such as fractional crystallization, crus-

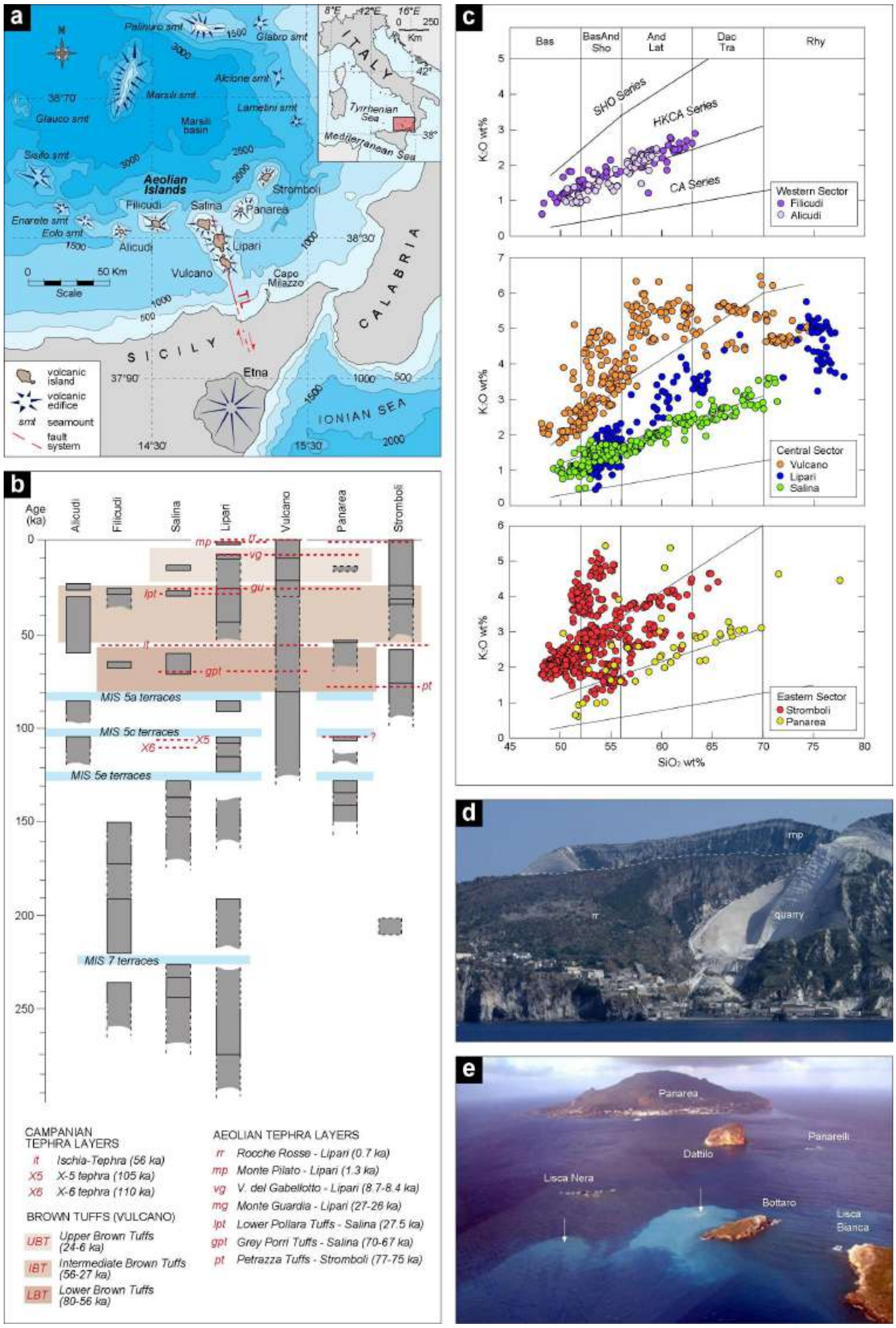


Fig. 23 - (a) Sketch bathymetry of the Aeolian Islands and seamounts, southern Italy (modified from Beccaluva et al., 1985). The trace of the Tindari-Letojanni fault system (TL) is shown. Depth contour lines in metres below sea level. (b) Schematic chronological framework of the Aeolian Islands volcanoes based on the available radiometric ages and correlation of marine terraces and tephra layers (see Lucchi et al., 2013a for age references). (c) SiO_2 vs. K_2O wt% (Peccerillo & Taylor, 1976) classification diagram of volcanic rocks from the Aeolian Islands volcanoes: CA=calcalkaline; HKCA=high potassium calcalkaline; SHO=shoshonitic; Bas=basalt; BasAnd=basaltic andesite; Sho=shoshonite; And=andesite; Lat=latite; Dac=dacite; Tra=trachyte; Rhy=rhyolite. Source of data: Alicudi (Lucchi et al., 2013d); Filicudi (Lucchi et al., 2013e); Salina (Lucchi et al., 2013b); Lipari (Forni et al., 2013); Vulcano (De Astis et al., 2013); Panarea (Lucchi et al., 2013c); Stromboli (Francalanci et al., 2013). (d) Panoramic view from the north of the Rocche Rosse obsidian lava flow (rr) effused from the NE side of the Monte Pilato pumice cone (mp), the flanks of which are intensely dissected by quarry activity. (e) Aerial view of the submarine gas bursts (arrows) occurred in 2002-2003 in the area of the minor islets; in the background the main island of Panarea is visible.

tal assimilation and mixing. The youngest activity in this area is recorded by the Canna neck and lava flows (29 ka), representing the few remnants of an independent volcanic center offshore the NW coast of Filicudi.

The island of Salina, in the center of the Aeolian archipelago, is the emerged portion of a large volcanic complex developed through time under the influence of the three regional tectonic trends acting in the archipelago (NNW-SSE, WNW-ESE, NE-SW). Six Eruptive Epochs between 244 and 15 ka produced lava flows, Strombolian scoriae and hydromagmatic and subplinian pyroclastic deposits with a CA basaltic to high-K CA dacitic composition (Lucchi et al., 2013b; Nicotra et al., 2014; Sulpizio et al., 2016), interrupted by major periods of quiescence, volcano-tectonic collapses and marine ingression phases during MIS 7 and MIS 5 stages (Lucchi et al., 2013b). The typical twin-volcanoes shape of the island is the result of the superimposition of the Monte dei Porri stratocone with the compound volcano formed by the nested Pizzo Capo, Monte Rivi and Monte Fossa delle Felci stratovolcanoes. Magma compositions and their variation through time are the result of contamination of primary magmas with the Calabro-Peloritano lower crust and subsequent differentiation dominated by polybaric fractional crystallization becoming dominant during the activity of Monte Fossa delle Felci and Monte dei Porri, whereas the older Pizzo Capo and Monte Rivi were fed by deep magma reservoirs located at the crust-mantle boundary (Nicotra et al., 2014). During the latest eruptive activities of the Pollara crater (27-15 ka) magma mixing and mingling processes triggered high-energetic basaltic to rhyolitic explosive eruptions.

The island of Panarea (Fig. 23d), together with the Basiluzzo islet, the small rocks of Le Formiche and the minor islets of Dattilo, Panarelli, Lisca Bianca, Bottaro, Lisca Nera are the remnants of CA to high-K CA andesite to dacite lava dome-fields, with minor pyroclastic products, representing the subaerial culminations of a large, mainly submerged, irregularly truncated cone-shaped volcanic complex. Volcanism on Panarea dates back to 155-149 ka, with different growth phases interrupted by erosional unconformities, whereas the latest explosive eruption occurred in the time range between 27-26 and 8.7 ka from an undefined eruptive vent in the area of minor islets (Calanchi et al., 1999; Lucchi et al., 2013c). Although long considered an extinct volcano going through a weak fumarolic phase of activity, Panarea was recently (2002-2003) characterized by signs of a possible eruptive renewal manifesting through spectacular submarine gas bursts occurred in 2002-2003 in the area of the minor Islets (Fig. 23d) and accompanied by seismic swarms and emission of high temperature fluids with magmatic geochemical signature (Caliro et al., 2004; Capaccioni et al., 2007).

The island of Lipari is the above-sea-level culmination of a broad, largely submerged volcanic complex belonging to the Vulcano-Lipari-Salina volcanic belt. This volcanic complex is composed of a large variety of lava flows, domes and coulees and pyroclastic successions produced by a spectrum of hydromagmatic, Strombolian to Subplinian and effusive activities, under the control of major regional tectonic trends and recurrent volcano-tectonic collapses. Volcanism has developed between c. 267 ka and the Medieval age, showing progressive West-East and South-North shifts of the eruptive vents through time, together with an overall chemical change of erupted products from early CA and high-K basaltic andesites to latest rhyolites with a notable gap in the dacites field and a steep increase in the K_2O content through time (Forni et al., 2013). The youngest volcanic eruptions produced the prominent Monte Pilato pumice cone (VIII century CE) (Fig. 23e) and the obsidian lava flows of Rocche Rosse and Forgia Vecchia (XIII century CE), sited in the north-eastern sector of the island, as well as highly dispersed white-coloured, fine-grained rhyolitic tephra layers (Pistolesi et al., 2021). At present, Lipari is in a quiescent stage with a few low-temperature fumaroles and hot springs.

4.4.2. Vulcano volcanic complex (130 ka-Present)

The island of Vulcano is the exposed summit of a largely submerged volcanic complex that is separated from the nearby Lipari only by a shallow-water and narrow strait of sea. Its development has occurred from about 130 ka to the well-known eruptive cycle of 1888-90 CE which gave origin to the definition of the “Vulcanian” eruptive style (Mercalli & Silvestri, 1891). Volcanism displayed a wide spectrum of eruptive-explosive activities from a number of volcanic edifices which become gradually younger moving north-westwards under control of tectonic structures associated to the NNW-SSE oriented Tindari-Letojanni fault system. The development of the two il Piano (ca.100 ka) and La Fossa multi-stage caldera structures (78-8 ka) have the same NNW-SSE elongation and display a progressive NNW-shifting through time (De Astis et al., 2013) (Fig. 24). Erupted products have variable K-alkaline character and degrees of evolution, ranging from basalts to rhyolites and changing through time parallel to the evolution of the volcanic system. It is noteworthy that successive large-scale hydromagmatic eruptions from the La Fossa Caldera emplaced the widely distributed Brown Tuffs (70-8 ka) pyroclastic products (Meschiari et al., 2020).

The Holocene activity mostly occurred from the La Fossa and Vulcanello eruptive centres (Fig. 24), fed by magmas ranging from shoshonites to rhyolites and alternating each other (De Astis et al., 2013; Nicotra et al., 2018).

The La Fossa stratocone (LFO) (Fig. 24a and b), in the middle of La Fossa caldera, was built during the last 5.5 ka through recurrent hydromagmatic to Vulcanian explosive eruptions, which gave rise to multiple dilute pyroclastic currents and fallout deposits, alternating with a few trachytic to rhyolitic lava flows. Compositionally, the LFO products reach the maximum degrees of magma evolution and alkali contents on Vulcano. The progressive growth of the cone occurred during different eruptive cycles interrupted by short intervals of quiescence. After the early activity (5.5-2.9 ka), when most of the present cone was built up to 250-300 m in elevation, a younger eruptive cycle during Roman to early Medieval times produced pyroclastic successions deposited by dilute to concentrated pyroclastic currents, two fallout layers and four lava flows from two different, intersecting craters (De Astis et al., 2013; Malaguti et al., 2022). The activity of the eccentric Forgia crater (Fig. 24c) in Medieval times is still not chronologically well constrained. Another explosive cycle of activity started from 1727 CE and culminat-

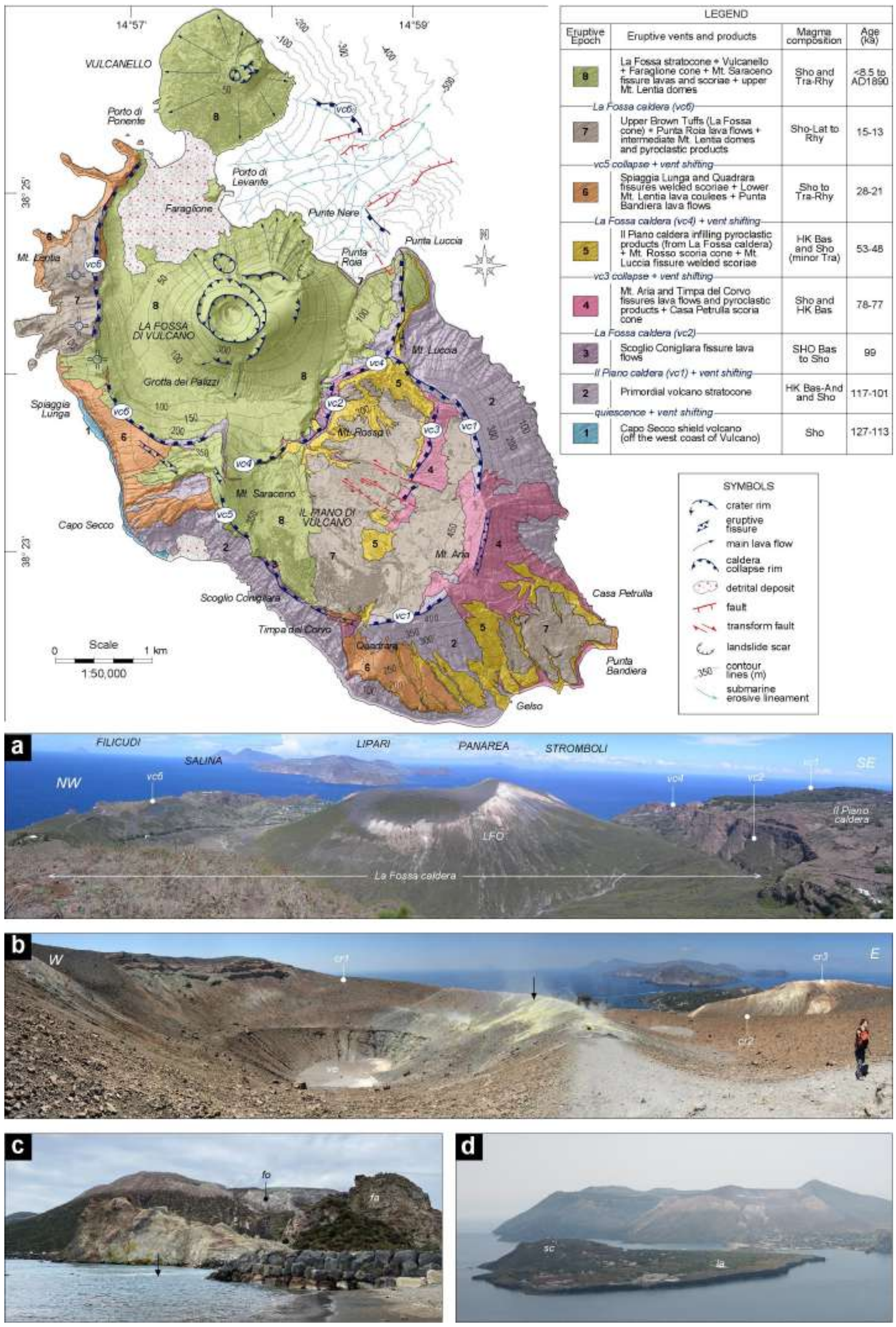


Fig. 24 - Simplified geological map of Vulcano merged on a shaded relief DEM of the island, showing the areal distribution of the erupted products and vents during its eruptive history, as well as the collapse rims corresponding to the multi-stage formation of the Il Piano and La Fossa calderas (modified from De Astis et al., 2013). Main landforms and caldera rims in the NE shallow-water submarine portion are also shown (from Casalbore et al., 2018). Age references from De Astis et al. (2013) and references therein. Magma compositions: basalt (Bas), andesite (And), shoshonite (Sho), latite (Lat), trachite (Tra), rhyolite (Rhy), high-K (HK), shoshonitic (SHO). (a) Panoramic view of the central sector of the island of Vulcano with the field evidence for the major volcano-tectonic collapse rims (vc1, vc2, vc4, vc6) leading to formation of the multi-stage il Piano and La Fossa calderas, and the La Fossa stratocone (LFO) standing out in the middle of La Fossa caldera. (b) View from the south of the summit area of LFO showing the crater (cr1) and vents (ve) of the AD 1888-1890 eruption, and its products characterized by bread-crust bombs and lava blocks; this crater is nested within older crater rims (cr2, cr3). The active fumarolic field (arrow) is visible along the northern rim of the AD 1888-1890 crater. (c) Outcrop of the intensely hydrothermally altered rocks of the Faraglione cone (fa) near to the Spiaggia di Levante beach, where intense gas emission in shallow-water areas (arrow) has occurred during the 2021-2022 unrest. In the background, the eccentric Forgia crater (fo) is visible along the northern flank of LFP. (d) View from Lipari of the Vulcanello lava platform (la) and nested scoria cones (sc), with the LFO in the background. All photographs credit F. Lucchi.

ed with the effusion of the outstanding Pietre Cotte rhyolitic lava flow in 1739 CE, the last effusive event from LFO (De Fiore, 1922). The Gran Cratere explosive eruptive cycle developed between 1739 CE and the latest paroxysmal activity of 1888-1890 CE, and was characterized by the distinctive emplacement of bread-crust bombs and lava blocks widely dispersed in the summit area of LFO (Fig. 24b). Since the last eruption, the LFO has shown only fumarolic activity, associated with shallow seismicity, fed by a shallow latitic magma body at a depth of 2.5-3.5 km (Paonita et al., 2013; Mandarano et al., 2016). Gas emissions are concentrated in several high-temperature active fumaroles along the northern rim of LFO crater, and subordinate points of emission and hot springs in the submarine areas near the Porto di Levante harbour (Granieri et al., 2006) (Fig. 24c). Unrest periods of increase in temperature above the usual range of 330-400 °C and in the magmatic component of the total gas flux (usually called ‘crises’) have occurred periodically (1920s, 1987-93, 1996-98, 2004-2009), with the last one started in September 2021 and still ongoing at the time of writing (Aiuppa et al., 2022).

Vulcanello is located along the northern border of La Fossa caldera and consists of three NE-SW-aligned, nested small Strombolian scoria cones from which a lava plateau, made up by a composite lava flow field of pahoehoe- to aa-lavas, formed (Fig. 24d). The oldest reported submarine to shallow-water eruption occurred during Roman times (II century BC, Stothers & Rampino, 1983), whereas most of the subaerial activity took place up to early Medieval times (De Astis et al., 2013; Malaguti et al., 2022), with latest eruptions around the XVI century and fumarolic activity persisting until 1878 CE (Keller, 1980). In the early stages, Vulcanello developed as an independent islet and was successively linked to the main island of Vulcano through the progressive accumulation of ash erupted from La Fossa forming the isthmus area between 1525 CE and 1550 CE (Barbano et al., 2017).

4.4.3. Stromboli composite volcano (85 ka-Present)

Stromboli is a largely submerged composite volcano well-known for its persistent “Strombolian” eruptive activity from historical times. Controlled by a regional NE-SW structural system, the exposed portion of Stromboli was entirely developed from ca. 85 ka to present times (e.g. Gillot and Keller, 1993; Risica et al., 2019), whereas the Strombolicchio neck is dated to ca. 204 ka. The successive epochs of activity, represented by Paleostromboli I-III (85-34 ka), Vancori (26-13 ka), Neostromboli (13-4 ka) and Recent Stromboli (<2.4 ka) (Fig. 25), mainly erupted mafic products (ca. <56 wt% SiO_2) covering a large compositional affinity from calc-alkaline to shoshonitic and potassic-alkaline series (Hornig-Kjarsgaard et al., 1993; Francalanci et al., 2013). These epochs were repeatedly interrupted by small, summit calderas and lateral collapses cutting the NW and SE flanks of the volcano down to the submerged basis of the edifice (Tibaldi, 2001; Romagnoli et al., 2009) (Fig. 25a, b). They largely conditioned the location of active vents and erupted products, with the current Sciara del Fuoco (SDF) scar developed during the last 13 ka. The activity has been mostly supplied from summit vents through alternating effusive and Strombolian eruptions, whilst more intense Subplinian explosions only characterized the Paleostromboli I and Vancori epochs, and subordinate (but notable) flank eruptions along the NE and SW buttressed flanks of the volcano took place during the Neostromboli one. Recurrent NW-dipping lateral collapses occurred during the Holocene in alternation with sequences of eruptions, and caused the unloading of the magmatic-hydrothermal system, likely triggering atypical hydromagmatic eruptions and pyroclastic density currents (PDC) that affected the lower slopes of the volcano (Lucchi et al., 2018). The Recent Stromboli epoch started during Roman times and resulted in the construction of the summit Pizzo scoria cone, whilst the San Bartolo fissure erupted pahoehoe lava flows that largely invaded the NE coastal sector forming a fan-shaped delta, and represented the last effusive activity outside the SDF. The present-day activity of Stromboli, starting from (at least) early Medieval times, has occurred from a crateric terrace located at elevations of c. 750 m a.s.l. near the SDF headwall (Rosi et al., 2013)(Fig. 25c). This activity is typically characterized by continuous active degassing (puffing) and rhythmic, discrete, weak to moderate explosions lasting a few seconds (Fig. 25d) that eject dark highly-porphyritic (hp) scoriaceous lapilli, bombs and ash, accumulating in the area around the craters and along the steep SDF scar. Occasional more violent explosions, covering a large intensity range from major explosions to paroxysms (Bevilacqua et al., 2020), produce convective plumes rising up to 7-8 km (Fig. 25e) and eject mingled scoriaceous hp and “pumiceous”, light-coloured and low-porphyritic lapilli, bombs and spatter, together with lithic blocks derived from deeper craterization of the upper conduit and vent system. Ballistic fallout can overcome the SDF walls impacting the flanks of the cone, and occasionally reaching the inhabited areas of Stromboli and Ginostra, whereas short-lived PDCs due to column collapse generally flow within the SDF (Fig. 25f). During historical times, paroxysms occurred repeatedly in the XVI-XVII and the XX centuries, with the more energetic event in 1930 and 1944, whilst recent paroxysms occurred on 5 April 2003, 15 March 2007, and the close events of 3 July and 28 August 2019 (Giordano & De Astis, 2021; Metrich et al., 2021; Ripepe et al., 2021; Viccaro et al., 2021). Effusive eruptions from summit craters or lateral vents/fissures opened at various elevations within SDF, recurrently interrupt the ‘normal’ Strombolian activity, lasting from ca. 1 day to a few months (e.g. in 2002-2003, 2007, 2014)(Fig. 25g). For (at least) the past three centuries, lava flows have descended the steep slopes of SDF, and eventually reached the sea (e.g. in 2007, 2014, 2022), representing a negligible hazard for people and infrastructures. Up to the Middle Age, lava flows have episodically surmounted the SDF lateral rims and expanded along the upper to lower flanks of the cone.

Recurrent gravitational instabilities and collapses at various scales along the oversteepened slopes of SDF are recorded during the last century (Barberi et al., 1993), eventually resulting in tsunami waves along the Stromboli and surrounding coasts, as observed associated to the December 2002 submarine and subaerial landslide developed along the SDF (Tinti et al., 2005).

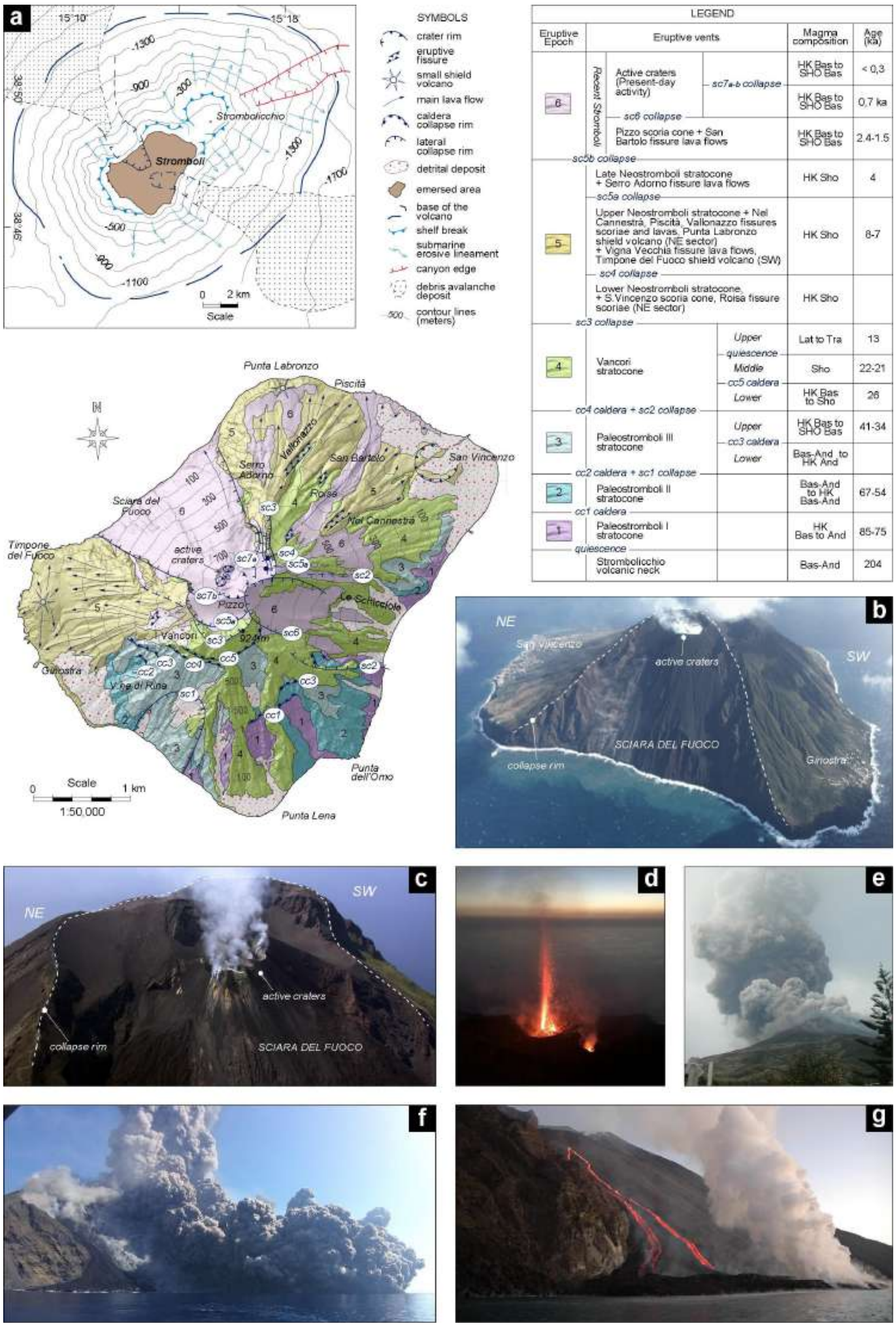


Fig. 25 - Simplified geologic map of Stromboli (modified from Francalanci et al., 2013) and morpho-structural sketch map (a) of its submerged flanks (modified from Romagnoli et al., 2013). Altitude points are in metres above sea level. The map is based on a DEM-shaded relief image courtesy of DICEA (Dipartimento di Ingegneria Civile Edile e Ambientale), Università di Roma La Sapienza (research project funded by the Italian Department of Civil Protection). Age references from Francalanci et al. (2013) and references therein. Magma compositions: basalt (Bas), basaltic-andesite (Bas-And), andesite (And) shoshonite (Sho), latite (Lat), trachyte (Tra), high-K (HK), shoshonitic (SHO). (b) Aerial view of the Stromboli composite volcano and the Sciara del Fuoco collapse scar cutting its NW flank, with the active craters visible near its headwall (Credit Italian Civil Protection). (c) Aerial view of the summit area of the Stromboli composite volcano showing the NE-SW aligned active craters near the headwall of the Sciara del Fuoco collapse scar (Credit INGV). (d) Strombolian activity of the summit craters (Credit M. Pistolesi). (e) Eruptive column formed during the 5 April 2003 paroxysm (Credit INGV). (f) Pyroclastic currents formed along the Sciara del Fuoco collapse during the 3rd July 2019 paroxysm (unknown author). (g) Pahoehoe lava flows forming a lava delta along the Sciara del Fuoco collapse during the 2007 effusive activity (Credit C. Romagnoli).

4.4. Mt. Etna (500 ka to Present)

(S.B., M.G., E.N., M.V.)

The eruptive history of Mt. Etna began about 500 ka (De Beni et al., 2011) in the eastern Sicily, an area of geodynamic complexities, where the Apennine-Maghrebian Chain overlaps the undeformed margin of the African continental plate, the Hyblean Foreland, bounding westwards the oceanic Ionian lithosphere (Lentini, 1982; Cristofolini et al., 1985; Branca et al., 2004; 2008; 2011a,b; Fig. 26).

Volcanic activity in the area developed after an early long period of scattered basaltic magmatism in the Hyblean Plateau that migrated northwards during Middle-Late Pleistocene towards the region of the present Mt. Etna (Neri et al., 2018). Volcanism in this area began with fissure-type eruptions with tholeiitic affinity emplacing lava flows in submarine and subaerial environment (Tanguy et al., 1978). Remnants of the submarine activity are pillow-lavas and columnar outcrops on the eastern shore north of Catania (about 500 ka; Gillot et al., 1994; Corsaro & Cristofolini, 1997; 2000; De Beni et al., 2011; Fig. 26). Subaerial tholeiites date back to 330 ka (De Beni et al., 2011) and mostly crop out on the SW slopes of the volcano (Fig. 26). The erupted lavas gradually changed in composition from subalkaline towards Na-alkaline, while the fissure activity mainly concentrated on the Ionian coast along the Timpe fault system starting from about 220 ka (Tanguy et al., 1997; Corsaro & Cristofolini, 1997; Branca et al., 2004; 2008; 2011a,b; Fig. 26).

During the last 130 ka the major eruptive axes migrated westwards, gradually aligning themselves on a central axis. This transition from fissure to central vent activity led to the construction of a series of central-conduit edifices producing Strombolian to Plinian eruptions in the area of the present-day Valle del Bove. As the central activity stabilized at ca. 57 ka (De Beni et al., 2011), a huge stratovolcano, the Ellittico, grew up. The intense eruptive activity both effusive and highly explosive of the Ellittico created a roughly 3600 m-high cone, which constitutes the frame of the present edifice of Mt. Etna (Fig. 26). The Ellittico activity ended 15 ka ago after four Plinian eruptions (Coltelli et al., 2000; Del Carlo et al., 2017) that created a 4 km-wide caldera. This became the site of persistent volcanic activity that progressively led to the emplacement of a new stratovolcano, the Mongibello, over the last 15 ka (Fig. 26). The formation of the Valle del Bove, a horseshoe depression about 7 x 4.5 km wide that is located on the eastern flank of Mt. Etna, dates back to the activity of the Mongibello stratovolcano, namely to 7478 - 7134 BCE (Malaguti et al., 2023).

The activity of Mongibello marks the beginning of the recent volcanic history of Mt. Etna, which also displayed a wide range of eruptive styles from purely effusive to violent explosive (Branca & Del Carlo, 2004). Historical records of eruptions date back to the Greek colonization of Sicily (734 BCE), passing through other chronicles during the Roman Age and the XVII century (Branca & Abate, 2019) with reference, respectively, to the 122 BCE Plinian eruption (Coltelli et al., 1998) and the impressive 1669 eruption (Corsaro et al., 1996; Nicotra & Viccaro, 2012; Kahl et al., 2017). The last few centuries were dominated by effusive and Strombolian eruptions (Branca & Del Carlo, 2005) with emission of hawaiitic products and subordinate mugearites (Viccaro & Cristofolini, 2008). Over the last five decades, the eruption frequency and explosivity increased, while the composition of volcanic products shifted towards K-trachybasalts (Tanguy et al., 1997; Tonarini et al., 2001; Viccaro et al., 2011; Viccaro & Zuccarello, 2017).

Mt. Etna has been particularly active during the last 30 years, experiencing continuous morphological modifications (Harris et al., 2011). Most of the morphological changes were caused by the opening of eruptive fissures on the flanks of the volcano that produced lava flows, as in 1991-1993, 2004-05 and 2008-09, and large scoria cones, as during the 2001 and 2002-2003 eruptions (Tonarini et al., 1995; Clocchiatti et al., 2004; Métrich et al., 2004; Andronico et al., 2005; Allard et al., 2006; Ferlito et al., 2012; Palano et al., 2017). Since nineties, there was also a clear growth in the number of paroxysmal eruptions from the summit craters (Fig. 26), which generated extensive eruptive columns, abundant tephra fallouts and ash-lapilli dispersion up to hundreds of km far from craters (Andronico et al., 2021). Main paroxysmal episodes occurred in short-term sequence of multiple events lasting weeks to years. Remarkable sequences were in 1998-1999 (23 events in 6 months), 2000 (64 events in 6 months) and 2007 (7 events in 7 months).

The last decade has also seen the development of some of the most energetic, yet long-lasting paroxysmal sequences of recent times, which have been fed by the most volatile-rich magmas ever found at Mt. Etna (Zuccarello et al., 2021). These include a three-year-long sequence of 44 lava fountains during 2011-2013 (Behncke et al., 2014; Giuffrida & Viccaro, 2017; Giuffrida et al., 2018; Zuccarello et al., 2022), and the very recent sequence of 62 events that took place between December 2020 and February 2022. These sequences alternate with periods of dominant effusive activity from both the summit and the volcano flanks (Viccaro et al., 2019; Borzì et al., 2020; Giuffrida et al., 2021), with the only brief interlude of two short, but powerful paroxysmal series in December 2015 (4 events) and May 2016 (3 episodes) (Corsaro et al., 2017; Calvari et al., 2018; Cannata et al., 2018; Zuccarello et al., 2022).

The recurrence of highly explosive events at the summit, which produce severe tephra fallout and episodic formation of pyroclastic density currents due to partial collapses of the South East Crater cone (Behncke, 2009a; Ferlito et al., 2010; Scollo et al., 2013; Costa et al., 2023), and effusive flank eruptions from vents located at low altitude (Del Negro et al., 2013) have the potential to cause significant socio-economic damages because of the high probability to impact the densely populated areas around the volcano, as well as hundreds of thousands of tourists who visit Mt. Etna yearly.

4.5. The Hyblean volcanism (M.G., M.V.)

The Hyblean volcanism began in Upper Triassic ca. 200 Ma (Cristofolini, 1966) and continued until Lower Pleistocene, with both submarine and subaerial episodes of mafic volcanism widely scattered across the area of southeastern Sicily (Carbone & Lentini, 1981; Schmincke et al., 1997). No volcanism is known to have occurred from the Upper Cretaceous to the Middle-Upper Miocene. The Mesozoic volcanism produced lavas of Na-alkaline affinity that outcrop in the southernmost and eastern part

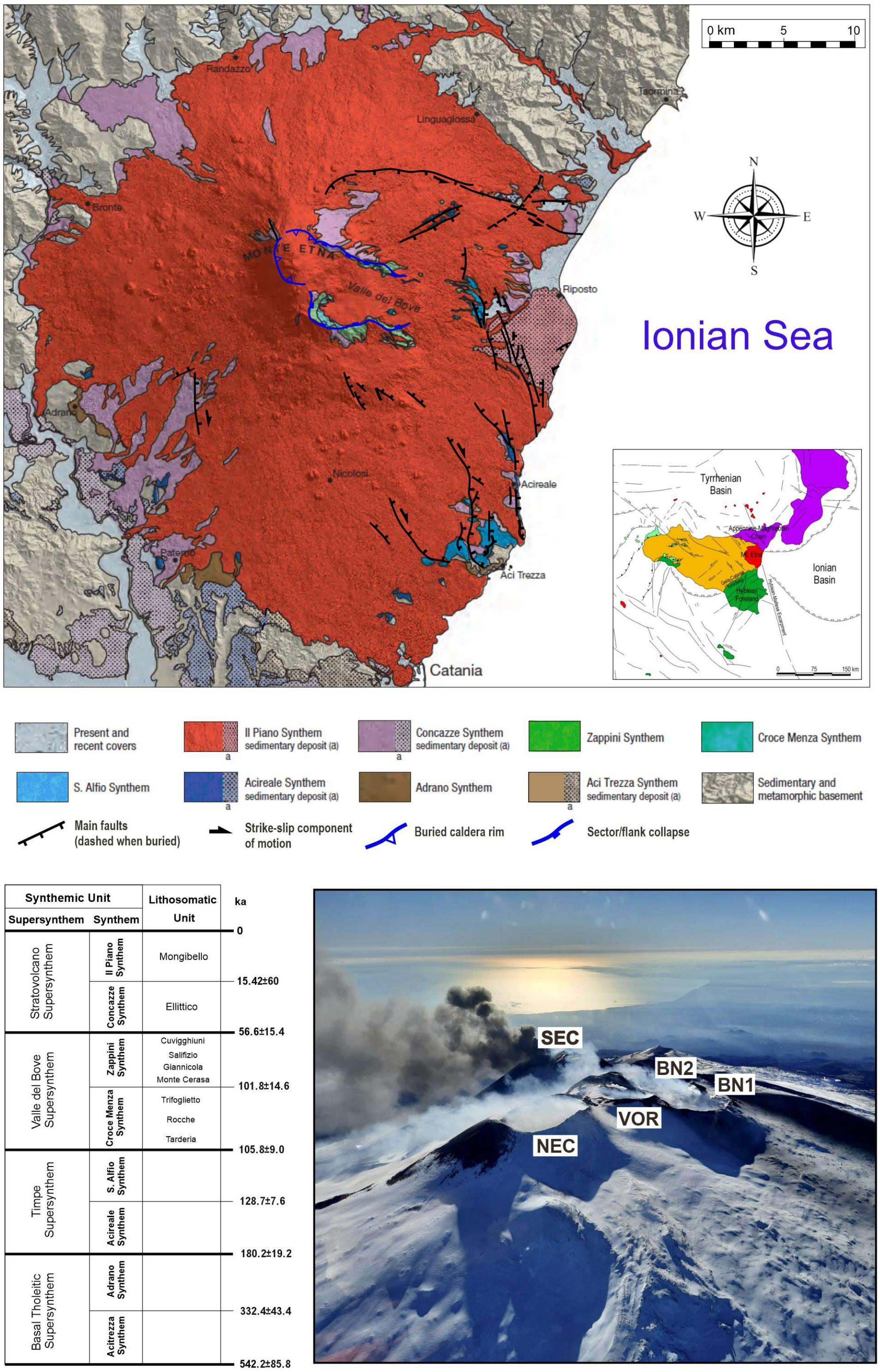


Fig. 26 - Synthemic geological map of Mt. Etna volcano with the main morphotectonic features (upper panel, modified after Branca et al., 2011); a scheme of the stratigraphic relationships of the volcanic succession using the supersynthem, synthem and lithosomes classifications and the radioisotopic ages have been also reported down-left in the figure (modified after Branca et al., 2011b) and aerial view of the Mt. Etna summit area taken on January 2021 with the 5 active craters, namely (lower panels): Voragine (VOR), North East Crater (NEC), Bocca Nuova 1 and Bocca Nuova 2 (BN1 and BN2), South East Crater (SEC).

of the Hyblean Plateau. A hiatus of ca. 50 Ma separates these earlier emissions from the later activity that resumed during Miocene in the central-northern area of the plateau with eruptions of both tholeiitic and alkaline magmas (De Rosa et al., 1991; Tonarini et al., 1996; Beccaluva et al., 1998; Trua et al., 1998). The Neogene-Quaternary volcanism includes two different eruptive cycles: a) the Miocenic cycle, prevalently alkaline in composition (mostly basanites and alkali-olivine basalts), which is dominated by explosive eruptions testified by diatreme-related volcanoclastic deposits bearing a significant amount of mantle xenoliths (Scribano, 1987a, b; Scribano et al., 2009; Suiting & Schmincke, 2010; 2012); b) the Plio-Pleistocene cycle almost exclusively characterized by relatively basic magmas, ranging in composition from extremely silica-undersaturated alkaline magmas to silica-oversaturated tholeiites. The Plio-Pleistocene cycle was the most widespread, taking place over an area of ca. 500 km^2 at the northern margin of the Hyblean Plateau. Large volumes of tholeiitic products were erupted in a shallow marine environment during the Upper Pliocene, followed by at least five episodes of alkali basaltic volcanism between the end of the Upper Pliocene and the Lower Pleistocene. The first of these episodes emplaced in the northernmost sector of the Hyblean Plateau a considerable volume of alkali basalts as pillow breccias, subaerial lavas and scoria deposits (Behncke, 2009b), while the latest eruptions (ca. 1.4 Ma; Schmincke et al., 1997) emitted scarce volumes of basanitic and nephelinitic lavas.

4.6. The Sicily Channel

(A.C., M.G., S.R., M.V.)

4.6.1. General background

The Sicily Channel is set in the thinned continental crust of the Pelagian block (Fig. 27a), a rifted zone in the Sicilian Maghrebian Chain foreland (Catalano et al., 2014). The main morphotectonic features are three deep tectonic troughs, NW-SE oriented, namely the Pantelleria, Linosa and Malta grabens, which exert control on magma generation and ascent (Fig. 27a). Magmatism in the Sicily Channel is widespread and results in the occurrence of several volcanic submarine centres (Civile et al., 2008; Rotolo et al., 2006) and two subaerial volcanic complexes, the Pantelleria (Fig. 27b) and Linosa islands. Submarine volcanism began around 10 Ma, with the latest event of the 'Ferdinandea' ephemeral islet emersion in 1831 CE and the 1891 CE 'Foestner' entirely submarine eruption (Conte et al., 2014; Coltelli et al., 2016; Kelly et al., 2014). Submarine volcanic rocks are mafic and dominantly mildly alkalic regarding the serial affinity (Rotolo et al., 2006; White et al., 2020). Ascent of mafic magmas is driven by tectonic structures, with the exception of Pantelleria, where the route to the surface is facilitated in the northern part of the island (Giuffrida et al., 2020) and inhibited in the central sector, with production of abundant felsic derivative products.

4.6.2. Pantelleria (320-4 ka)

The subaerial volcanic activity at Pantelleria started around 320 ka and is characterized by emission of pantelleritic products emplaced as lava flows and pumice fallout deposits (Civetta et al., 1984; Mahood & Hildreth, 1986; Rotolo et al., 2013; Jordan et al., 2018). Nine ignimbrite eruptions covered the island (entirely to partly) in the age interval 181-46 ka and alternated with inter-ignimbrite periods, during which tens of local eruptive centers were active (Jordan et al., 2018, Rotolo et al., 2021; Fig. 27c). Two ignimbrites out of nine, namely the Capre and Green Tuff Formations (age 140 and 46 ka, respectively) are associated with caldera collapses and related scarp remnants (La Vecchia and Cinque Denti caldera, respectively; Fig. 27b, Speranza et al., 2012; Rotolo et al., 2013). Other five potential caldera collapses (now buried) have been hypothesized based on the occurrence of thick pyroclastic breccia horizons within or at the top of some ignimbritic deposits (Jordan et al., 2018). Over the period 181-46 ka, basaltic manifestations were rare and limited only to some lava flow units dating back to 118 ka and 83 ka (Civetta et al., 1984) that were found in the northwestern part of the island (Fig. 27c). The post-Green Tuff volcanic activity was dominated by low-energy Strombolian eruptions and effusive events, mostly inside and along the rim of the Cinque Denti caldera. This activity produced pumice fall deposits, lava flows and domes of dominant pantelleritic compositions (Civetta et al., 1988; Rotolo et al., 2007). An updated stratigraphy of the post-Green Tuff evolution is given in Rotolo et al. (2021) based on high resolution $^{40}Ar/^{39}Ar$ ages (Scaillet et al., 2011) and paleomagnetic data (Speranza et al., 2010), with two periods of activity after the Green Tuff eruption, instead of the five proposed by Civetta et al. (1988).

Basaltic volcanism accounts for ca. 5 % of outcropping rocks and is centered in the NW sector of the island with short lava flows and scoria cones (Giuffrida et al., 2020). Mafic volcanism is dated (K/Ar) in the age interval 31±13 ka and 27±10 ka, with possible younger ages close to 10 ka (unpublished data; Civetta et al., 1984; Fig. 27d).

4.6.3. Linosa (1.06-0.5 Ma)

Linosa is a small volcanic island (ca. 6 km^2) whose subaerial portions represent only 4% of the entire volcanic edifice, which extends for ca. 20 km along the NW-SE tectonic lineament characterizing the whole Sicily Channel (Fig. 28; Romagnoli et al., 2020). While no age data are available for its submarine activity, the subaerial eruptive history can be constrained between ca. 1.06 and 0.5 Ma, during which volcanism was effusive to explosive, mostly hydromagmatic, and occurred over three major periods (Rossi et al., 1996). Erupted magmas show limited compositional variation, being only mafic (alkali basalts to hawaiites/mugearites), slightly more primitive than the Pantelleria basalts. Rare intrusive lithics in pyroclastic products have more evolved, syenitic compositions never erupted as magma.

4.7. Seamounts of the Tyrrhenian Sea (6 Ma-Present)

(R.D.R., S.d.V., P.D., F.L., E.N., R.S., C.A.T., M.V.)

The Tyrrhenian Sea started its opening 15 Ma during the Apennine orogenesis as a result of back-arc extensional processes. In the southern sector, the roll-back of the Ionian slab caused the SE migration of the Calabrian Arc and accelerated the process of back-arc extension, leading to the progressive thinning and substitution of the continental crust with a new oceanic one. This process is responsible for the opening of the Vavilov (6

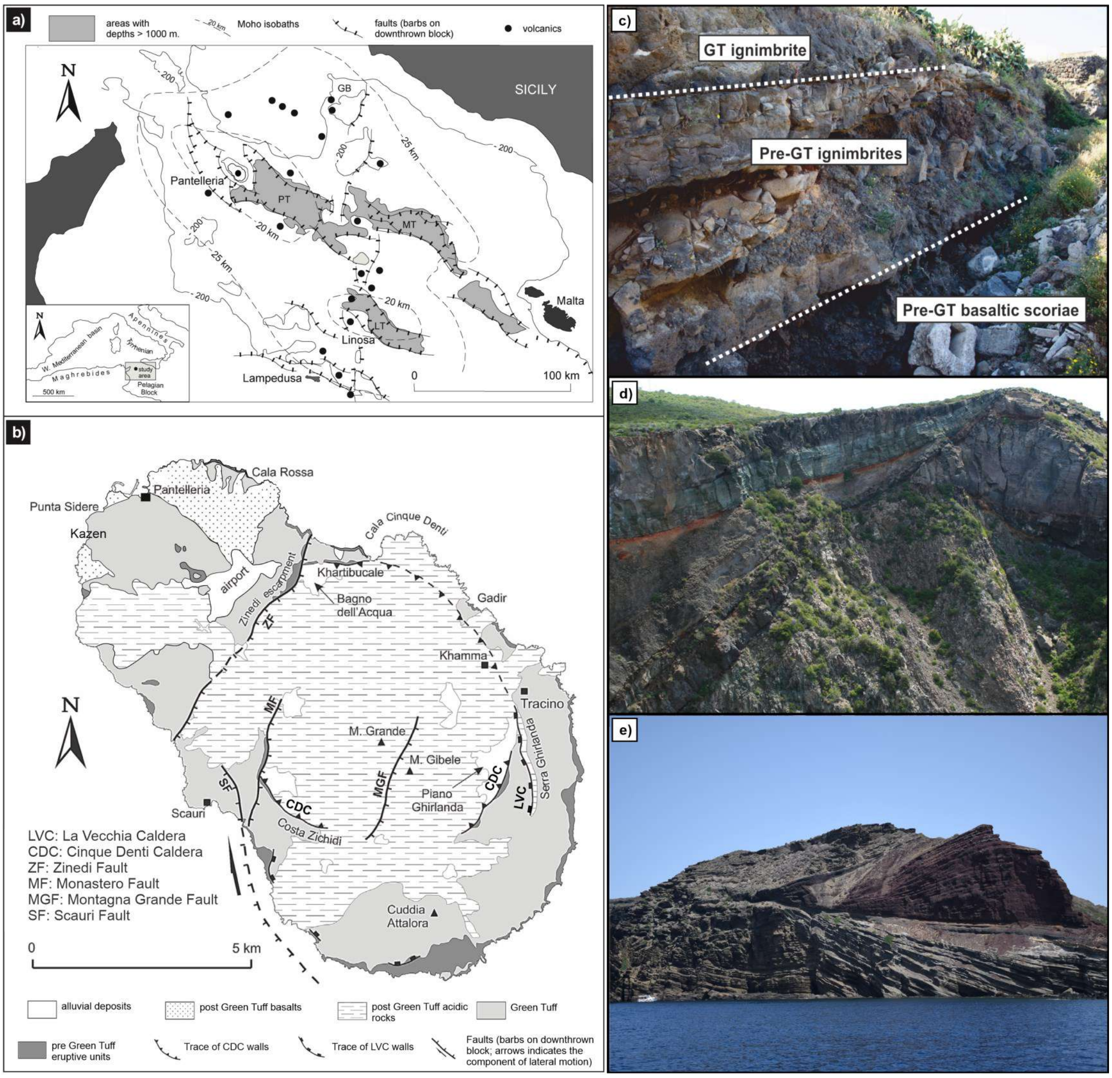


Fig. 27 - (a) Schematic structural map of the Sicily Channel (PT: Pantelleria through; MT: Malta through; LT: Linosa through, GB: Graham Bank) and (b) Geological and structural sketch map of the island of Pantelleria (modified after Catalano et al., 2014); (c) volcanic succession in the area of Kazen at Pantelleria: pre-Green Tuff scoriaceous basaltic rocks are visible at the bottom of the succession, which is constituted in the upper portion by other pre-Green Tuff ignimbrites and finally by the Green Tuff ignimbrite; (d) Pantelleria Island (Salto La Vecchia): La Vecchia Caldera scarp. The caldera wall dips 50° NW (left) and separates flat-lying pre-caldera ignimbrites (to the right of the caldera wall), with concordant post-caldera ignimbrites and one fallout deposit. The sequence is topped by the Green Tuff ignimbrite (age 46 ka), the last of the nine Pantelleria ignimbrites; (e) Linosa island (Punta Calcarella): hydromagmatic and fallout deposits of three major volcanic phases.

or 4.5 - 2.5 Ma) and Marsili (1.8 Ma-pres.) oceanic basins and the birth and growth of a great number of seamounts (Fig. 28a; Trua et al., 2004), whose ages decrease from NW toward SE.

Marsili is a huge (70x30 km) volcanic submarine complex, elongated along a N-S direction with its summit placed 3 km above the seafloor. Although the oldest products date back to 1 Ma, Marsili is considered an active volcanic system due to: i) seismic activity typical of volcanic and hydrothermal areas (D'Alessandro et al., 2012); ii) volcanic gases detected above the water column (Lupton et al., 2011); iii) the products of the two eruptive episodes occurred 3000 and 5000 years ago, recognized at its top (Iezzi et al., 2014). A temporal progression characterizes the Marsili magmatic activity from IAB-like basalts with calc-alkaline affinity to OIB-like basalts (Trua et al., 2007). The summit area is marked by volcanic cones perfectly preserved, with some evidence of lava flow fields and volcano-tectonic structures (Ventura et al., 2013). Magnaghi and Vavilov are the other two important seamounts of the Tyrrhenian Sea, with an elongated shape over 20 km. Magnaghi seamount is related to the ancient activity (3-2.7 Ma; Serri et al., 2001), whereas Vavilov seamount formed during the Quaternary (0.73-0.1 Ma) and it could still be considered active (Savelli & Ligi, 2017). Although they lie on a seafloor with MORB-like geochemical signature (Trua et al., 2007), the few rocks collected at Magnaghi and Vavilov seamounts show a mildly alkaline OIB-like geochemical affinity (Trua et al., 2007). Palinuro seamount, consid-

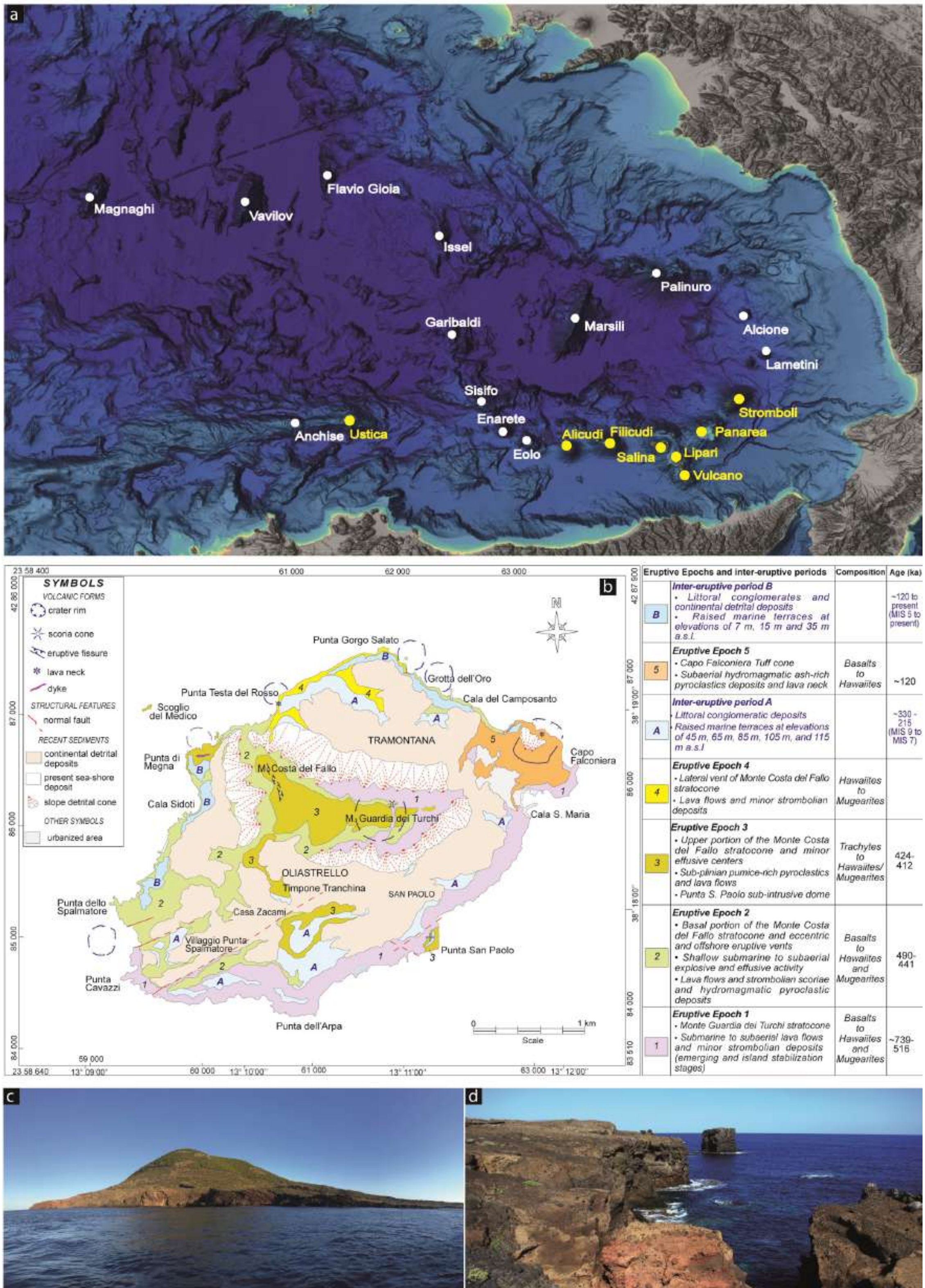


Fig. 28 - (a) Morphometric map of the Tyrrhenian Sea; seamounts are labeled in white, whereas emerged volcanoes in yellow; (b) Simplified geological map of Ustica; (c) Panoramic view of the western coast of the Ustica island and of the Monte Costa del Fallo stratocone (Eruptive Epoch 2); (d) Panoramic view of the northern seacliff of the Ustica island and of lavas and scoriae belonging to Eruptive Epoch 3.

ered active or quiescent (Milano et al., 2012), presents a 50 km E-W elongated shape, resulting from the superimposition of different volcanic edifices. The composition of emitted products is very similar to the emerged portion of the Aeolian Islands (Trua et al., 2004).

4.8. Ustica (730-ca.130 ka)

(R.D.R., S.d.V., P.D., F.L., E.N., R.S., C.A.T., M.V.)

The island of Ustica is the exposed summit of a largely submerged NE-SW elongated volcanic complex rising about 2000 m above the Tyrrhenian seafloor, including an extensive field of submerged volcanic cones (up to thirty) to the north of the island. It is located over a transitional zone between a northern domain characterized by thinned oceanic crust (ca. 8 km thick) and a variably thick (25-30 km) western-southern domain of continental crust (Savelli, 1988; Sulli, 2000).

Based on literature data and new unpublished studies, the eruptive history of the island of Ustica, entirely developed during the Quaternary (de Vita et al., 1998; Bonomo & Ricci, 2010), is arranged into five epochs of activity (Fig. 28b) that emplaced submarine to subaerial volcanic rocks with a Na-alkaline signature. The first epoch (730-516 ka) includes submarine basaltic lavas emitted from fissural vents and basaltic to hawaiitic and mugearitic lavas relevant to the large Mt. Guardia dei Turchi stratocone, which progressively emerged from the sea. Basaltic to hawaiitic and mugearitic lava flows and pyroclastic products of the second epoch (490-441 ka) are related to the Monte Costa del Fallo stratocone and its lateral eruptive vents, partly located offshore the west coast of the island. The third epoch (424-412 ka) is characterized by the emplacement of hawaiite/mugearite lava flows (Fig. 28c and d) and trachytic pumice pyroclastic products from the Monte Costa del Fallo stratocone, together with the products of a series of lateral vents including the trachytic, amphibole-bearing sub-intrusive body exposed at Punta San Paolo (Alletti et al., 2005). The fourth eruptive epoch was fed by lateral dykes in the NW sector of Monte Costa del Fallo stratocone that emplaced hawaiitic to mugearitic lava flows and minor strombolian deposits. Finally, the fifth eruptive epoch includes the alkali-basaltic tuff cone and lava neck of Capo Falconiera, which was active at about 130 ka (de Vita et al., 1998). The whole island of Ustica has been eroded and reworked by different orders of marine terraces, whereas the volcanic products of Capo Falconiera are interlayered within marine terraces of the Last Interglacial (de Vita et al., 1998).

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