Spatial and temporal volcanotectonic evolution of Santorini volcano, Greece

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Abstract
Volcanic and tectonic activities in the Aegean region have controlled the evolution of Santorini volcano, including changes in the shape and size of the island through time. Previous studies associate much of the island’s volcanic activity with the presence of regional faults, but a comprehensive volcanotectonic study that clarifies the relationship between dyking and faulting in the island has not been made. Here we present a detailed structural analysis focused on the northern caldera wall of Santorini, where numerous dykes and faults outcrop and can be studied in the mesoscale. To augment our discussion of dyke and fault interactions, we combine previous volcanological and geophysical observations with our structural analysis to report the volcanotectonic evolution of the northern part of the island and design a conceptual spatial-temporal model. We mapped 91 dyke segments and 15 faults and classified the latter, where possible, with respect to their observed or recorded kinematics, their size, and the active stress field under which they were formed based on prior geophysical data. We relate our observations to a mechanical unconformity within the northern caldera wall. Our field observations, coupled with previous numerical, geophysical, and volcanological studies, offer insights on the interaction between dykes and faults and indicate the conditions under which the faults facilitated magma emplacement, or not, during the volcano’s activity. Our analysis attempts to answer an essential question: under what conditions do crustal faults facilitate or inhibit magma propagation to the surface, with application to the island of Santorini.

Keywords Santorini volcano · Volcanotectonics · Dykes · Faults · Structural analysis · Mechanical unconformity

Introduction
Dyke propagation to the surface is the primary process that feeds volcanic eruptions (e.g. Rubin and Pollard 1987; Tibaldi 1992, 2015; Gudmundsson 2011, 2020; Acocella 2021). Dykes normally make their own paths, but some become deflected into pre-existing fractures such as faults or joints (e.g. Delaney et al. 1986; Gudmundsson 1999, 2011, 2020; Browning and Gudmundsson 2015; Ebinger et al. 2008; Le Corvec et al. 2018; Thiele et al. 2021; Drymoni et al. 2021). Hence, field observations of dyke-fault relationships are essential for understanding dyke-path formation and the likelihood of a volcanic eruption (Gudmundsson 2020).

At Santorini, it is suggested that magma propagation occurs predominantly within two principal fault zones, the Kameni and Kolumbo fault zones (Mountrakis et al. 1996; Druitt et al. 1999) and the Christiana-Santorini-Kolumbo rift zone (hereafter CSK) (Nomikou et al. 2013; Hooft et al. 2017; Heath et al. 2019). During the 2011–2012 unrest episode, magma accumulation beneath the Kameni fault zone further demonstrated the existence of a seismogenic crustal fault (Newman et al. 2012; Feuillet 2013). Although, in this case, the fault did not facilitate magma movement, various sections of the fault did slip seismically, partly due to crustal stress transfer associated with the inflating magma chamber
(Parks et al. 2012; Lagios et al. 2013; Papadimitriou et al. 2015), but the crustal magma chamber did not rupture or produce a dyke (Browning et al. 2015). This episode demonstrated that changes in the local stress field may occur without dyke injection — probably a very common situation in active volcanoes. In Santorini, the relation between changes in the local stress field and magma propagation remains somewhat elusive. In particular, the interaction between dykes and faults needs to be investigated from the point of view of the geological record and for reliable future dyke propagation forecasting.

In this paper, we study magma transport, primarily through dykes, in Santorini volcano. We present new field observations and structural analysis combined with previous geophysical studies, seismic tomography, and numerical models to improve our understanding in magma transport and reconstruct the spatial and temporal volcanotectonic evolution of the Santorini volcano. The focus is on the interaction between dykes, faults, and the associated stress fields active over time. More specifically, we address the question: Do the faults facilitate or inhibit the transport of magma in the shallow part of the crust? Here we investigate the interaction of dykes with the observed northern caldera wall faults to determine their role in the emplacement of magma and further their role in constructing the various morphologies present at the volcano’s lifespan.

Geodynamic setting and the Aegean microplate

Active deformation in the Aegean is predominantly associated with two primary processes: (1) the relative motion of the Anatolian strike-slip fault zones and (2) the northward subduction and rollback of the African plate underneath the Eurasian plate (McKenzie 1972; Le Pichon and Angelier 1979; Taymaz et al. 2007). The two Anatolian fault zones represent two intracontinental transform fault systems with (a) a dextral (North Anatolian fault system-NAFS) and (b) a sinistral (East Anatolian fault zone) motion, respectively (Taymaz et al. 2007; Bulkan et al. 2020). The dextral motion of the NAFS has played a significant role in controlling Aegean geodynamics. Following the common view (Barka 1992), the NAFS originated in the mid-Miocene as a structure that propagated gradually from the east (at 13 Ma) to the west (at 5 Ma). Nuriel et al. (2019) instead proposed that the NAFS started as the evolution of a pre-existing extensional normal fault system (at 42 Ma) that was reactivated and transferred to a dextral strike-slip fault zone (at around 11 Ma).

The dextral strike-slip faulting of the NAFS and regional subduction processes generated the crustal deformation that accommodated the movements along the strike-slip, normal, reverse fault systems (Taymaz et al. 1991; Benetatos et al. 2004; Taymaz et al. 2007; Nocquet 2012) and the volcanism in the Aegean region. The main structures can be summarised from the northern to the southern part of the Aegean as follows (Royden 1993; Kokkalas and Aydin 2013):

1. Right lateral strike-slip displacement due to the relative motion of the Anatolian block from the E to the W (and slightly to the WSW) at a rate of 20–25 mm/year
2. N-S extension due to slab rollback, 6–10 mm/year, in the back-arc (central Aegean continental crust), accommodated crustal thinning by as much as 20 mm/year since the Miocene. The thinning accelerated to 40–50 mm/year in the Quaternary, and so in the present day, the region experiences both a fast slab retreat (Brun et al. 2017).
3. E-W active crustal shortening due to gravitational collapse of the thickened crust into the mantle before the Miocene (Gautier et al. 1999; Jolivet et al. 2008) and SW-SSW differential slab retreat due to rapid migration of the Crete trench’s subduction hinge rather than towards Cyprus-Anatolia (Doglioni et al. 2002, 2007; Agostini et al. 2010).
4. Because of greater velocity of the Aegean microplate, it overrides the African plate faster than they converge resulting in the initiation of subduction (Doglioni et al. 2007). At a rate of 2–4 mm/year near to the subduction trench, localised lateral extension is found, which was due, in part, to the curved shape of the trench.

Active extension in the southern Aegean region occurred in pulses since the Miocene (Le Pichon and Angelier 1979; Jolivet et al. 2013) and was recorded by the following structural events: (1) the fast exhumation of high-pressure metamorphic rocks channelled through detachment faults between 24 and 15 Ma (Jolivet et al. 1996; van Hinsbergen and Meulenkamp 2006) and (2) graben formation along Plio-Quaternary high-angle faults due to crustal stretching (Taymaz et al. 2007). Those extension pulses also controlled the location of volcanism in the Aegean Sea (Fytikas et al. 1984; Papazachos and Panagiotopoulos 1993). Christiana Islands, Santorini, and the Kolumbo submarine volcanoes are developed on the faulted southern margin of Anhydros Basin along an NW-SE high angle rift zone, surrounded by graben and horst formations (Papazachos and Panagiotopoulos 1993; Konstantinou 2010; Nomikou et al. 2016; Hooft et al. 2017) (Fig. 1).

Geological framework

The volcanic complex of Santorini (Thira) is composed of five islands: Thira, Thirasia, Nea Kameni, Palea Kameni, and Aspronisi. Nea Kameni and Palea Kameni Islands (hereafter, named Kameni Islands) represents the latest products of the volcanic activity (Fig. 1c). They were formed inside...
a flooded caldera which was formed after the catastrophic Minoan eruption (at 3.6 ka) (Druitt et al. 1999), but the northern part of the caldera was already exposed at this time and since the pre-Minoan era (Athanassas et al. 2016).

At the north-east outer part of the Santorini complex lies the Kolumbo submarine volcano. The latter is formed atop an elongated strike-slip fault zone, the Kolumbo rift, along with at least 25 smaller volcanic cones which last erupted violently in 1650 AD (Nomikou et al. 2013, 2016, 2019).

The geology of the northern caldera wall and the deposits’ age were described in detail by Druitt et al. (1999) (and references therein) and Druitt (2015). Most of the deposits of the northern caldera wall were formed as part of the Peristeria stratovolcano (530–430 ka). The first explosive cycle (224–172 ka) in the studied area is represented by (1) ra, extrusions of rhyodacites at the contact between MV and MPI in NE Thera (~224 ± 5 ka); (2) ap2 (Cape Therma 3), pink siller facies of ignimbrite and red spatter agglomerate with abundant plutonic nodules (196 ka); (3) rp3 (Lower Pumice 2), rhyodacitic pumice (172 ka), the second explosive cycle (145–3.6 ka) (Figs. 1a and 2b). The second explosive cycle is represented by

(1) ap4 (middle tuffs — Megalo Vouno (hereafter MV) and Kokkino Vouno (hereafter KV) cinder cones and Kolumbo tuff ring), thick scoria-fall deposits of phenocryst poor andesite (54 ± 3 ka);
(2) as2 (Skaros lava shield), andesites and basalts of Cape Skaros (70–54 ka);
(3) ap5 (Upper scoria 2), andesitic thick spatter agglomerate (54 ± 3 ka);
(4) ao (andesites of Oia), andesitic lavas;
(5) rp6 (Cape Riva eruption), rhyodacitic pumice fall phase overlain by two ignimbrites (welded and non-welded) topped by a lithic-rich lag breccia (21.8 ± 0.4 ka); and
(6) rp7 (Minoan eruption), rhyodacitic pumice fall, overlain by surge deposits, phreatomagmatic ash topped by an ignimbrite interstratified with a lithic co-ignimbrite and lag breccias (3.6 ka).
Volcanotectonics of Santorini

The caldera wall’s recent morphology was formed by a circumferential ring fault that was reactivated during the last caldera collapse episode (3.6 ka). The ring fault forms an elliptical closed loop (11 km × 7 km in diameter) at the periphery of the caldera rim. The fault dips subvertically, and its displacement is at least several hundred meters.

Two volcanotectonic lineaments strike roughly NE-SW, namely the Kolumbo (KL-VT) and the Kameni (KM-VT) lines (Fig. 1c), and dissect the island of Santorini. Both are fault zones and represent the surface expressions of either basement strike-slip or normal faults (Druitt et al. 1999). The Kolumbo volcanotectonic (KL-VT) line (N 40°–50° E) is situated in the northern part of Thira. It is partly overlapped in the north-eastern part of the island by the Kolumbo rift zone (CSK) (Hübscher et al. 2015; Nomikou et al. 2016) (Fig. 1c). Onshore, it cuts the cinder cones of MV and KV and the Cape Kolumbos tuff ring (Heiken and McCoy 1984; Druitt et al. 1999) (Fig. 2). Recently, Heath et al. (2019) suggested an extrapolation of the CSK fault zone to the southwest up to a NE-SW striking dyke at the southern part of Thirasia Island. Their results proposed magma localisation along this NE-SW lineament. While it is possible, a dyke does not require a fault zone to form or propagate. In fact, the dykes in the northern caldera wall follow a range of strike orientations indicating that many, perhaps most, did not use the fault zone lineament (Drymoni et al. 2020).

There are four NE-SW trending normal faults (Druitt et al. 1999) between the mountains of MV and Mikros Profitis Ilias (hereafter MPI) along sections of the northern caldera wall (Fig. 2). Two of these faults were suggested to have been triggered by the Cape Riva eruption (rp6) and the Thirasia-Skaros complex’s collapse around 22 ka. The collapse accommodated the lateral movement of a large section of the edifice into the collapsing depression forming a small NE-SW striking graben in a section of the caldera wall (Druitt et al. 1999), which has been recently debated from geophysical studies (Tzanis et al. 2020).

Fig. 2 (a1) Geological map of the area modified from Druitt et al. (1999) showing the location of the dyke swarm from Drymoni et al. (2020). (a2) Illustration showing the location and motion of F1, F2, F3, F4 faults modified from Druitt et al. (1999), and the studied fault structures (F2, F3, F4, F5, F6, F7, F8, F9, F10, F11, F12, F13, F14, F15) during this survey. (b) Panorama of the north-east corner of the caldera wall showing the lithologies of the heterogeneous and anisotropic host rock from Druitt et al. (1999) as follows: Peristeria stratovolcano; (av1-3) basaltic to dacitic lavas, tuffs, scoria and breccia. First explosive cycle: (ra-ap2-rp3) rhyodacitic lavas and a pyroclastic density current atop a rhyodacitic pumice. Second explosive cycle: (ap4-as2) thick scoria-fall deposits under andesitic and basaltic lavas and finally the mapped fault structures and the emplaced dykes according to this study. (c) A closer look on the fault structures (F10, F11, F12) at the northern caldera wall formed in the av3 lavas of the old Peristeria edifice.
Earlier studies (Mountrakis et al. 1996) suggested that the MPI faults were simply part of a primarily dextral (N 35–40° E) striking strike-slip fault zone and a normal (N 40–70° E) striking fault system with subsidiary N-S trends. Those results were derived from measurements of the strike orientation of the dykes and faults collected along the caldera wall and based on the hypothesis that the dykes were exclusively emplaced along faults. It was from this deduction that the same authors defined a NNW-SSE stress orientation (σ3). However, recent field observations and analysis of the dyke paths at Santorini (Drymoni et al. 2020), as well as fracture-mechanics results (Broek 1982; Atkinson and Meredith 1987; Gudmundsson 2011), suggest this may not be the case since most dykes generate their own paths regardless of the distribution of crustal faults (Gudmundsson 2020). Thus, the dyke segments observed on the northern caldera wall are generally unlikely to have propagated along faults.

Druitt et al. (1999) proposed that three of Santorini’s calderas formed during two explosive silicic eruptions and at least partly using the Kameni and the Kolumbo fault zones to accommodate the collapse as well as to channel magma to the surface. The evidence for this is that each collapse event exhumed a set of previously existing cliffs and unconformities along the same scarp, except for the uppermost subaerial products which formed later.

The distribution of active faults and their evolution

The oldest fault system in Santorini is described by the presence of an E-W listric normal fault trend (Papanikolaou et al. 1988) located in the Christiana and Folegandros basins but was rotated gradually through time to become ESE-WNW trending (Perissoratis 1995) and was later cut by a series of N-S transfer faults (Piper and Perissoratis 2003) forming the Santorini-Ios ridge in the northernmost part of the island. The latest trends primarily affected the metamorphic basement (Atticocycladic massif) and the lower parts of the terrigenous sediments deposited in Late Pliocene as reported by seismic tomographic analyses (Perissoratis 1995). The initiation of Santorini’s volcanic activity since the Late Pliocene-Early Pleistocene is represented by the early centres of Akrotiri (650 ka) and could be associated with the transition of the E-W trending to NE-SW strike-slip faults (Perissoratis 1995). This transition developed between the Late Pliocene and Early Quaternary (Piper et al. 2007) and delineated the formation of the Kameni line.

In the mid-Quaternary, the previous E-W trend was replaced by NE-SW strike-slip faults indicating either the collision of the African plate or the buoyancy effects of the thicker continental crust (Bohnhoff et al. 2006; Piper and Perissoratis 2003). The continuous collision activated ENE-WSW trending faults (Pe-piper et al. 2005) and rotated the N-S active transfer faults to become NNW-SSE oriented (Piper and Perissoratis 2003). The first faults in the volcano were dextral strike-slip (Bohnhoff et al. 2006) and were initiated at about 200 ka (Piper and Perissoratis 2003). Subsidence of at least 600 m was reported to have begun during the first explosive cycle, especially in the Late Pleistocene (in the last 200 ka) (Piper and Perissoratis 2003).

N65° E strike-slip sinistral faults were initiated between 0.2 and 0.6 Ma at the Amorgos region, forming a basin at the Santorini-Anafi region during the Late Pliocene and a positive (horst) region during the Early Pleistocene (Piper et al. 2007). The latter could be associated with the sinistral strike-slip motion of the Anatolian block from the East (ten Veen and Meijer 1998).

The upper parts of the offshore deposits around Santorini, according to tomographic studies, belong to Quaternary volcaniclastic sediments characterised by two sets of normal fault trends (1) NE-SW (Bohnhoff et al. 2006) and (2) N65–75° E (Piper and Perissoratis 2003). They have been initiated as regional sinistral strike-slip faults and remained active during the Early Quaternary. The latest NE-SW trend is parallel to the Kameni VT line and rift zone that showed signs of deep seismic activity in the past (Fytikas et al. 1990) and remains active (Dimitriadis et al. 2009; Kolaitis 2011). The evolution of the faults at the northern caldera wall, as described above, is shown in Fig. 3.

Methods

We performed several field campaigns on the northern caldera wall and mapped the dyke swarm, the associated host rock, the observed volcanotectonic features, and the fault structures. We studied the fault segments by collecting data related to their location, attitude, sense of slip (if possible), and their interactions with the mapped dyke segments. Our focus was on the outcrop scale or mesoscale, that is, the scale of fractures as seen in the field, which ranges from tens of centimetres to tens of kilometres.

As almost all the observed faults dip in excess of 60°, we used the following criteria to define the kinematics of the studied faults (where possible): (1) striated surfaces (slickenlines or lineations) or bedding surfaces, (2) conjugate faults which indicate slip parallel to the fault-dip line, (3) drag folds, (4) syntectonic listric normal faults or filled fractures, (5) other fracture geometries that accommodate secondary crystallisation (hydrothermal veins, joints), (6) stylolites and Riedel shears, (7) marker horizons, and (8) stratigraphic host rock relations.

We also compared our collected data and volcanological and structural analysis with previous volcanological (Druitt et al. 1999), seismological (Dimitriadis et al. 2009; Kolaitis 2011), geophysical (Tzanis et al. 2020), structural
Mountrakis et al. 1996), and numerical (Drymoni et al. 2020, 2021) studies to design a conceptual model of the evolution of the northern caldera wall since the formation of Peristeria stratovolcano (at 530 ka).

**Field results**

**Faults**

The fieldwork campaigns revealed two size classes of fault zones that we broadly classify initially depending on the size of the displacement. Those fault zones in suite 1 accommodated displacements in excess of 10 m, as discerned from stratigraphic relationships but could not be directly measured from marker horizons. The fault zones in suite 2 accommodated displacements of less than 10 m and could be measured from marker horizons. In Fig. 2, we present the locations of the faults mapped by Druitt et al. (1999) (F_1, F_2, F_3, F_4) and those mapped during this study (F_2, F_3 and F_5–F_15).

The field observations revealed that suite 1 contained mainly normal with some of them to have strike-slip component which marked the boundary of a graben. Unit offsets on either side of the graben indicate at least 100 m of displacement. Because of limited exposure, the along-strike extent of the faults could not be measured, and slip indicators were also not easily observed. Other studies report the same faults as normal or strike-slip faults at the MPI locality (Mountrakis et al. 1996) or structures formed due to block tilting (Tzanis et al. 2020). The profiles and analyses of these faults are now considered and presented in Figs. 2 and 4 while their interpreted structural analysis can be found in Table 1 and is explained below.

Fault F_2 has an attitude of (dip/strike) 60–65°/015° and cuts the complete stratigraphy of the northern caldera wall. The uppermost layer of the footwall sequence (Ska-rros lavas — as2) was deposited between 70 and 54 ka (Druitt et al. 1999). It can be classified as a normal fault with an offset of at least 200 m which is based on the observed stratigraphic relationships. At least two dykes of varying composition and hence probably different populations were deflected into the fault. Remnants of the first explosive cycle (ra-ap2-rp3) are found only at the NE corner of the caldera wall (Fig. 2b). These products are cut by fault F_2 and are restricted only at the hanging wall. All horizons are horizontal except the bottom layer (ra) which dips to the NW by 40° (Fig. 4a). This implies that the unit “ra” was either deposited normally inclined at the slopes of the pre-existing summit or that the F_2 fault was at some point a listric normal fault and was activated again after the deposition of “ra” (around 224 ka). However, no other kinematic indicators were observed. Later reactivations could have occurred due to the deflection of different dyke populations with either a normal sense or a strike-slip motion (from focal mechanisms, Dimitriadis et al. 2009); hence, we classify this fault F_2 as a reactivated normal fault but with a modern strike-slip motion (Table 1; Fig. 4a).
F₃ fault has an attitude of 65°/030° and cuts the whole length of the stratigraphy restricted to old lavas (Peristeria stratovolcano, 530–430 ka) at the MPI locality (Fig. 2). It can be classified as a normal fault with an offset of at least 70 m which is based on the observed stratigraphic relationships. On the bottom of the fault zone, we observed signs of SW dipping of the stratigraphy and displacement (Druitt et al. 1999) (Fig. 4b) and a recent strike-slip movement based on recent geophysical studies (Hübscher et al. 2015; Tzanis et al. 2020). Several dyke segments were deflected into the fault, and one of them is found at the top of MPI. The dykes were of similar colour and thickness, did not crosscut, and hence, it was difficult to discern relative ages (Table 1; Fig. 4b).

F₂ and F₁ form a graben structure as detailed by Druitt et al. (1999) based on the stratigraphic relationship of the collapsed block. Our field studies agree that the structure stratigraphically resembles a graben structure as the middle part is depressed; however, the faults do not dip towards each other but instead dip towards the same direction (SE) which forms a system of normal faults. Both structures (a graben or a system of normal faults) can result scarcely tilted, so the current motion (Tzanis et al. 2020) can be explained regardless.

F₄ fault is classified as a strike-slip fault based on the presence of negative flower structures which are structural indications of transtension/strike-slip faulting. It has an attitude of 90°/020° and hosts a 1.5-m-thick mafic dyke which became deflected into the fault lower in the sequence. The dyke has similarities in chemistry with the Skaros deposits (Drymoni 2020), and if we assume they are part of the same sequence, this indicates it was active during the formation of the Skaros shield (Drymoni et al. 2021).

F₈ is classified as a normal fault according to the observed marker horizons; it has an attitude of 65°/015° and strikes parallel to F₂, but it does not cut any of the Skaros lavas (70–54 ka) and hence could not have been active for as long as F₂, but it must be older than 70 ka (Figs. 2b and 4). The minimum offset is 50 m, and there are no dyke segments observed cutting or deflecting into the fault zone (Table 1; Fig. 4a).

F₉ has a similar attitude to both F₂ and F₇. The stratigraphic horizons, through which the fault cuts, show a normal sense of displacement (approximately 10 m offset), and its activity is restricted to the Peristeria stratovolcano. No dyke segments cut or become deflected into it (Table 1; Fig. 4a).

F₁₁ may have originated as a strike-slip fault and later or simultaneously acted as a normal fault or vice versa. Strike-slip displacement is evidenced through a negative flower structure in the uppermost part of the fault while markers of normal displacement were observed from the surrounding stratigraphic horizons. No dykes cut or become deflected into it (Table 1; Figs. 2 and 4b).

F₁₂ is restricted in the Peristeria stratovolcano, and our field analysis classified it also as a strike-slip fault showing positive flower structures and drag folds. No dykes were found to cut or become deflected into the fault (Table 1; Figs. 3 and 4b). Faults F₁₁–₁₅ are likely to be strike-slip faults, but again, no dykes interacted with these faults (Table 1; Fig. 4b).

F₁ and F₄ faults are reported as normal faults (Druitt et al. 1999); we do not expand on this interpretation since these fault zones were several hundreds of meters away from the dyke swarms and the main area of interest.

Some of those faults with offsets smaller than 10 m have been previously reported (Mountrakis et al. 1996; Druitt et al. 1999), but our structural analysis revealed more intricate features, as shown in Fig. 4 and summarised below as follows:

F₆ has an attitude of 65°/345° and permeates only the av1 layer of Peristeria stratovolcano (530–430 ka) which is composed of lavas, scoria, tuff, and hyaloclastites. It has a vertical length of at least 20 m and an offset of 4–5
Table 1 The structural analysis of the mapped faults at the northern caldera wall

| Fault (this study) | Attitude (dip/strike) | Observed Kinematic indicators (this study) | Age of activity (this study) | Activity after 1964 (from Dimitriadis et al. 2009; Kolaitis 2011) | Dyke deflection (this study) | Data supported by previous studies |
|-------------------|-----------------------|---------------------------------------------|-----------------------------|-----------------------------------------------------------------|-------------------------------|-----------------------------------|
| F_2               | 60°-65°/015°          | Normal: 200 m min offset calculated from stratigraphic relationships Peristeria — present | Active                      | At least 2 dyke populations (mafic and felsic)                   |                               | Normal faulting: Mountrakis et al. (1996) (structural analysis), Druitt et al. (1999) (volcanological analysis) Strike-slip faulting: Mountrakis et al. (1996) (structural analysis), Dimitriadis et al. (2009) (focal mechanisms), Tzanis et al. (2020) (geophysical analysis) |
| F_3               | 65°/030°              | Normal: 70 m min offset calculated from stratigraphic relationships After ap4 (cinder cones) — present | Inactive                    | 1 population (at least 3 dykes)                                |                               | Normal: Mountrakis et al. (1996) (structural analysis), Druitt et al. (1999) (volcanological analysis) Strike-slip: Mountrakis et al. (1996) (structural analysis), Tzanis et al. (2020) (geophysical analysis) |
| F_5               | 90°/020°              | Strike-slip: negative flower structures Skaros — present | Active                      | 1 dyke                                                          |                               | Strike-slip faulting: Mountrakis et al. (1996) (structural analysis) |
| F_6               | 65°/345°              | Normal: 3 m min offset calculated from marker horizons Av1 (Peristeria volcano) | Inactive                    | 1 dyke and 1 inclined sheet                                      | N/A                           |
| F_7 (double fault) | 45°-55°/015°         | Normal: 4–5 m min offset calculated from marker horizons Av1 (Peristeria volcano) | Inactive                    | 3 dykes                                                         | Normal faulting: Mountrakis et al. (1996) (structural analysis) |
| F_8               | 65°/015°              | Normal: 50 m min offset calculated from stratigraphic relationships After 54 ka (Skaros shield) | Inactive                    | Not found                                                       | N/A                           |
| F_9               | 65°/015°              | Normal: 10 m min offset calculated from stratigraphic relationships After 54 ka (Skaros shield) | Inactive                    | Not found                                                       | N/A                           |
| F_10              | 60°-70°/010°          | Zone of small-scale faults (shear fractures) After 54 ka (Skaros shield) | Inactive                    | Not found                                                       | N/A                           |
| F_11              | 90°/040°              | Normal: 10 m min offset calculated from stratigraphic relationships Strike-slip: negative flower structures After 54 ka (Skaros shield) | Inactive                    | Not found                                                       | N/A                           |
m which is recorded by a clear marker horizon in a scoria unit. We observed multiple dyke injections that became deflected into the fault zone (Table 1; Fig. 4a); hence, we characterise it also as a reactivated one. Since this fault dissects only the lowest part of the avi Peristeria unit, the dykes were emplaced during the Peristeria and/or Skaros subswarms.

F7 has an attitude of 45–55°/015°. We observed two 10-m dip faults (a normal and a reverse fault) with convoluted crosscutting geometries and a minimum offset of 3 m. Two dykes were deflected in total into both faults (Table 1; Fig. 4a), but it is unclear if they belonged to the same population or not.

F10 is a zone of small-scale faults (shear fractures) which have an attitude of 60–70°/010°. No dykes were found to become deflected into this structure (Table 1; Fig. 4b).

The kinematic analysis of the northern caldera wall faults and their various kinematic expressions (normal, reverse, strike-slip) demonstrate the dynamic volcanotectonic evolution of the area through time. Only five of the fifteen faults studied (F2, F3, F5, F6, F7) showed signs of dyke deflection (Fig. 5). Also, only 12 dyke segments out of the 91 (13%) demonstrated an interaction with a fault. Similarly, only two of those five faults (F2, F6) demonstrate multiple dyke deflections during possibly different dyke emplacement episodes and hence magmatic reactivation indications. F2 records the deflection of two different dyke populations (both mafic and felsic) which indicates that the fault was active during their emplacement while F6 records the deflection of one composite dyke followed by a later emplaced inclined sheet.

![Fig. 5 3D Schematic illustration showing the volcanotectonic setting of the NE part of the northern caldera wall during the present day. The depressed block between the F2 and F3 faults is now tilted as only F3 is active. Stereographic projections of the dykes and faults are also given](image-url)
The volcanotectonics of the northern caldera wall and the Peristeria unconformity

An angular unconformity of approximately 4 km in length interrupts the Peristeria lava units and forms a 30-ka hiatus in between the silicic-andesitic (av2) and andesitic-basaltic (av3) lava flows, an observation also reported in previous studies (Druitt et al. 1999). An unconformity may be generated by non-volcanic processes such as erosion, sea-cliff formation, or large-scale landslides. However, none of these processes can be easily identified here since the process that formed the unconformity occurred during the earliest volcanism and has been subsequently overprinted by later volcanotectonic events. It is then reasonable to associate the unconformity with volcanic processes such as caldera collapse, eruptive vent migration (Lucchi et al. 2013), or phreatomagmatic eruptions (Geshi and Oikawa 2008). Lucchi et al. (2013) suggest two genetic end-members for unconformity forming processes in volcanic environments (i) allogenic unconformities (large-scale processes) caused by global-scale phenomena such as sea level fluctuations and (ii) autogenic unconformities (medium-scale processes) that are primarily generated by internal modifications in a volcano such as (1) continued volcanic inactivity, (2) rapid locality change in eruptive vents, and/or (3) volcanotectonic events (caldera or lateral collapse).

During this study, we mapped the intermittent exposure of the observed unconformity, and we studied the dyke emplacement within its vicinity. At the Epta-Paides cape locality (Fig. 6), dyke segments were found to become arrested before or at the unconformity contact while others clearly dissect it. This is likely because the unconformity acted as a discontinuity (i.e. a unit with a contrasting Young’s modulus) close to its formation which could make it a stress barrier and promote dyke arrest. With time, the discontinuity became mechanically stiffer; hence, later populations (e.g. the Skaros subswarm) were able to propagate and feed surface eruptions and lava flows (Fig. 6). Here, the location and size of the unconformity (sharp 4-km long contact), the intermittent exposure of the contact, the funnel-shaped silicic fossilised lavas on the northern caldera wall which could possible relate to a fossilised vent, and the existence of another already recorded unconformity driven by caldera collapse (Druitt et al. 1999), close to the study area, could imply that the angular unconformity reported here could be also possibly related to a caldera collapse mechanism.

Druitt et al. (1999) mapped three occurrences of sporadically distributed av2 lava deposits where the first one formed a triangular funnel-shaped feature below MV summit and suggested that it is the remnant of a lava-filled vent or lavas plastered against an ancient steep wall. The other two represent a silicic andesitic dome at the angular unconformity and a massive silicic andesitic flow at the base of MPI (Fig. 2, a1). We offer an alternative mechanical emplacement of some of these deposits as instead relating to possibly columnar joint emplacement as shown in Figs. 7 and 8. We traced a light yellowish colour lava layer along the panorama. We noted a spatial and textural similarity to a suite of subvertical silicic dykes of the same profile. The lower, upper, and tops of the lavas are undulating. They are characterised by well-developed columnar joints and are much thicker than the other lava flows that exhibit structures associated with slow cooling. All the dykes macroscopically are all yellowish to light-coloured, and XRF analyses on two of them have shown more felsic compositions than the rest of the dykes (Drymoni 2020). They are seen to strike radially from the Peristeria summit and have a variable thickness between 1.1 and 8 m. In Figure 8, it can be seen that their localities genuinely overlap with the more silicic lavas above. This suggests that some of the silicic dykes may have fed the av2 eruption. In contrast, others became deflected due to the dissimilar layering of the caldera wall (Drymoni et al. 2020).

Discussion

Indications of volcanotectonic processes during the evolution of Peristeria stratovolcano

Despite the evidence of connections between the earliest formed collapse calderas and fault zones at Santorini (Druitt et al. 1999), it remains unclear how the Peristeria stratovolcano relates to the most recent caldera collapses. In the central part of the island’s caldera wall, at Fira, Druitt et al.
(1999) demonstrated evidence of unconformities related to caldera-forming collapse events. In the northern part of the island, there exists also evidence of at least one large angular unconformity, but its origin is not well-constrained. This lack of constraint is primarily because the caldera wall in this section is comparatively old and complex. In particular, all of the previous four caldera collapse events have overlapped the area and destroyed the potential corroborating evidence. As such, any field interpretation related to the unconformity in the northern caldera wall will tend to be speculative. Hence, although the existence of the angular unconformity maybe related to volcanotectonic processes, such as a caldera collapse, the connection of the unconformity with non-volcanic erosional or mass-wasting processes such as landslides, surface or wave erosion, and gullying cannot be ruled out. Here we provide four lines of evidence that support a caldera-collapse origin of the northern caldera wall angular unconformity.

**Location and size of the unconformity**

The unconformity is restricted to the northern part of the caldera wall and is thus lacking in the north-eastern region (MPI locality) where the av2-av3 contact is normal and no hiatus is
observed. This discrepancy between the northern and northeastern parts of the caldera indicates tectonic control since, in the case of surface or wave erosion or gullying, the unconformity could be expected to occur along the entire contact. We also consider the unconformity, which can be traced for 4 km, as being of a size that could fits with its being of volcanic origin based on the classification by Lucchi et al. (2013). In their study, sea-level change fluctuations form large-scale unconformities that may be traced across entire islands (perhaps for tens of kilometres). By contrast, the present unconformity is restricted only to a part of the island. These observations may be regarded as supporting origin through volcanotectonic processes.

**Intermittent exposure of the contact**

The unconformity along the caldera wall does not delineate the av2-av3 contact as would be expected if the unconformity was the result of erosion (Schwartz et al. 2018). Instead, most of the unconformity contact is represented by the av1-av3 contact. The av2 deposits are found sporadically in a range of thicknesses (1–100 m) along the unconformity, and as shown by this study (Fig. 8), the exposure localities of the av2 lavas, and thus the av2-av3 contact, are set by yellow-coloured dykes. No dyke-fed eruptions are observed at the north-eastern part of the caldera, and there is a continuous av2 deposit of around 100–140 m thick. We suggest that the volume difference of the av2 deposit is such that it cannot be explained as erosional surfaces, a landslide or sea-cliff formation, particularly in lava flows. Thus, the large volume of av2 lavas on the north-eastern part of the caldera wall, which do not occur on the northern part of the caldera wall, could also possibly be related to a caldera collapse during the formation of the Peristeria volcano.

**Morphology of the silicic andesitic lavas**

The northern caldera wall is characterised by a funnel-shaped exposure of an av2 lava which locally has a maximum thickness of 100 m. This lava has been interpreted as either the remnant of a lava-filled vent or lava flows plastered against an ancient steep wall (Druitt et al. 1999). Our field observations define a separate area that possibly forms another large av2 deposit (Fig. 8f). The latter lies atop a silicic dyke, the thickest dyke found in the caldera cliffs (8–10 m). Such a dike might be expected to feed large volume silicic andesitic lavas in the caldera wall.

**The existence of another well-constrained unconformity, which sets the wall of a previous caldera**

This is a scenario that has been evidenced in Santorini by Druitt and Francaviglia (1992) and suggests that an unconformity related to caldera collapse has been already reported on the island. Scoria cones are not only commonly fed by volcanic fissures eruptions but are also commonly located at the periphery of a ring fault (Gudmundsson 2011). They are then either fed by a ring dyke or by inclined sheets or thin radial dykes that have been deflected into the ring fault (Browning and Gudmundsson 2015; Saunders 2001). The location of the two later calderas (at 21 ka and 3.6 ka), according to previous studies (Druitt et al. 1999; Athanassas et al. 2016), overlaps with the trace of the cinder cones and indicates that the later collapses occurred at least partially along the previous ring faults. We did not find field evidence to support a connection of the cinder cones with basement faults, as reported by Druitt et al. (1999), and geochemical analyses of the cinder cone samples do not show absolute similarities (Drymoni 2020). As a result, if the scoria cones are fed by deflected dykes at the periphery of a ring fault, then a caldera collapse (forming ring faults) could have preceded the dyke emplacement.

Based on the evidence and suggestions above, we propose that a piston-like collapse generated by a dip-slip ring fault could have been a likely mechanism to explain the unconformity, a process that has been already proposed for later caldera collapse events (Roche and Druitt 2001; Roche et al. 2000). Experimental caldera collapse models, which explore asymmetric subsidence scenarios (Lavallée et al. 2004), indicate that such an event could have tilted the MPI block and induced slumping towards the area of the maximum subsidence since the MPI stratigraphy shows greater vertical motion. However, incremental collapse has also been suggested for basaltic volcanoes (Michon et al. 2011). In both cases, it is likely that displacement along the faults occurred in many slip events rather than in one large displacement event, as evidenced from the stratigraphic correlations and age of the deposits below the cinder cones of MV, KV, and MPI localities — and well documented in the collapse calderas in Fernandina (Galapagos Islands) in 1968 (Filson et al. 1973) and in Kilauea in 2018 (Neal et al. 2019). Our interpretation of stratigraphic and volcanotectonic evolution of the northern caldera wall is shown in Fig. 9.

**Timing of faulting and dyke emplacement — synthesis**

Peristeria volcano was formed between 530 and 430 ka (Druitt et al. 1999), and we propose that was possibly partially destroyed by a collapse caldera, dated from an angular unconformity at the av2/av3 contact, at around 496 ka. Both the F₆ and F₇ normal faults cut only the av1 unit and exhibit interactions with dykes that likely emplaced during the formation of the Peristeria subswarm. We make this assumption because the dykes were not emplaced higher than the Peristeria succession. At approximately the same time, a series of crustal strike-slip and normal faults trends
were activated and associated with two important geodynamic events in the region. The transition of the E-W to a NE-SW fault trend at around 650-ka initiated volcanism on Santorini and transferred it to the northern part of the island around 100,000 years later, creating the first major rift zone of the island (Perissoratis 1995; Piper et al. 2007). We suggest that the F2 fault was active during this period of volcanism at Peristeria volcano and accommodated predominantly normal motions that partly formed a horst and graben-like structure detailed by Druitt et al. (1999). The same tectonic setting explains the reverse motion of a second F6 fault zone (Fig. 4a). The F2 normal fault was also likely active during a dyking event under the Peristeria volcano (Fig. 10a) as felsic (yellow) dykes were found to be deflected into the fault zone (Fig. 2a). Still, we did not find evidence of shearing along the dyke walls. The only silicic subaerial products that could have been fed by this dyke swarm were thus of the av2 unit. Those faults did not accommodate dyke emplacement, so they were unlikely to be active during the Peristeria volcanism. There may have been a triggering of an N65° E sinistral strike-slip system (0.2–0.6 ka) associated with the large-scale motion of the Anatolian block from the East (ten Veen and Meijer 1998; Piper et al. 2007), which corresponds to the early stages of the Santorini-Amorgos basin. In both cases, the faults were likely subsidiary faults to a larger ENE-WSW trending system. However, as most of the eruptive activity during this period was concentrated in the central part of the volcanic complex, it is reasonable to assume that most of the dyke injections that controlled the volcanism were also restricted to either the central part of the complex or as fissure eruptions on the slopes of the volcano. Hence, the dykes observed in the northern part of the island predominantly fed flank eruptions (Fig. 10). There was no synchronicity between the activity on the faults and the emplacement of dykes during the first explosive cycle. This implies that the dykes deflected into the F2 normal fault could not have been coeval with the eruptive events of the first explosive cycle.

A collapse caldera formed around 180 ka during the first explosive cycle (360–172 ka), and the eruptive activity then later migrated west to create the Simandiri edifice (Druitt et al. 1999). The Simandiri dyke swarm did not interact with any of the northern caldera wall fault zones as it is far from the fault segments. Additionally, there were no deposited
lavas flows associated with the Simandiri edifice found at the eastern part of the island (Fig. 10b).

During the middle Pleistocene, around 126 ka, the first NE-SW normal and strike-slip faults were triggered in the area and coincide with the formation of cinder cones at MV, KV, and the Kolumbo tuff ring (125–80 ka) eruptions. The scoria cones are composed of tephra developed from explosive eruptions of mafic and intermediate magmas (ap4). However, we found no evidence of scoria feeding dykes within faults, and hence, we assume that those dark-coloured mafic dykes that deflected into the \( F_2 \) normal fault zone belong to a later subswarm.

During the last 200 ka years, there have been around 600 m of subsidence offshore of the main Santorini complex.
(Piper and Perissoratis 2003). Such rapid subsidence is possible in narrow grabens (i.e. those with less than 20 km width). The primary mechanism is caused by horizontal extension and crustal thinning associated with dyke emplacement in areas where faulting is already active (Jarvis 1984). If active extension in the Southern Aegean occurred in pulses since the Miocene, we propose that the F2 and F3 normal faults were active earlier than the 70-ka caldera reactivation. Thus, the bulk of the subsidence could instead be associated with the 180-ka collapse caldera formed during the 1st eruptive cycle or simply by the general extension in the Amorgos basin that activated large-scale normal faults. We suggest that both the F2 and F3 normal faults correspond and were formed during the subsidence within the area (Fig. 10c).

During the second major explosive cycle, forming the Skaros and Thirasia shields, two sets of normal and strike-slip faults striking N40-55° E and N65-75° E were also formed (Piper and Perissoratis 2003; Bohnhoff et al. 2006). We suggest that the F2 and F3 normal faults were active during the formation of both shields, which is evidenced by several dykes deflecting into sections of both faults (Fig. 2). While we note two normal faults (F1 and F3) that dip parallel to each other, the stratigraphic correlations and focal mechanisms imply both subsidence and strike-slip motion; hence, we expand the interpretation to suggest that they originated as normal faults but later accommodated strike-slip motions. The F3 strike-slip fault was formed during this explosive cycle as a mafic dyke used the fault for part of its path.Geochemical analyses have shown that this dyke has similarities with the Skaros lavas (Drymoni 2020), and hence, we denote that the F3 strike-slip fault has been activated since the Skaros shield formed (70–54 ka). However, the F3 fault did not accommodate slip again after the second explosive cycle and according to current seismological data, it is no longer seismically active.

Graben subsidence coeval with the 21-ka Cape Riva caldera formation has been interpreted as a failed breach of the caldera which displaced a large block towards the Skaros shield summit (Druitt et al. 1999). Geochemical correlations between the host rock and the dykes emplaced at the MPI locality exhibit similarities with the Peristeria, Skaros, and Thirasia subswarms (Drymoni 2020). This implies that the “graben” was still active during the most recent dyke emplacements, i.e. during the formation of the Skaros shield (70–54 ka) and hence possibly much earlier than the 21-ka age, which was proposed by Druitt et al. (1999). We attribute the stratigraphic correlations of the graben to both near-constant activity of the F2 and F3 faults, as reported by observations of dyke deflection into the faults (Drymoni et al. 2021) and a subsidence driven event (Druitt et al. 1999). However, if the F3 fault is still inactive now, then the block no longer accommodates subsidence. This agrees with recent geophysical studies (Tzanis et al. 2020), which propose tilting rather than subsidence of the area. Finally, between the F2 and F3 faults, a series of minor normal (F8 and F9), small-scale faults (F10), and strike-slip faults (F11 and F12) were formed. Dyking was synchronous to faulting for only 12 of the 91 mapped dyke segments, and none of these dykes interacted with any of the faults F5 through to F15. This suggests that all these faults became active after the volcanic activity in this part of the island had ceased (Fig. 10d).

In the present day, the northern caldera wall forms a scarp that traces both the 21 ka and 3.6 ka ring faults. The caldera faults, and the majority of the faults observed in the caldera wall, are seismically inactive, but faults F2 and F5 have shown historical signs of activity related to strike-slip motion according to seismological catalogues since 600 B.C. (Makropoulos and Burton 1985; Makropoulos et al. 1989; Papazachos and Papazachou 1997; Papazachos et al. 2006; Dimitriadis et al. 2009; Kolaitis 2011) (Fig 10e).

Conclusions

Mapping of the northern caldera wall at Santorini volcano allowed us to record 15 faults which we classified based on the amount of offset they accommodated (meters or offsets of tens of meters up to 100 m), as well as the type of displacement (normal, reverse, strike-slip) and the active stress field in which they formed (e.g. extensional, transtensional). For the first time at Santorini, we have coupled results from field studies with previous results from numerical models and geophysical studies/seismic tomography to characterise the interaction between dykes and faults. Combining these methods allowed us to reconstruct the geological history of the northern part of Santorini and consider the dynamic active volcanotectonic setting that formed the morphology of the current island part (Fig. 10).

The main question of this paper concerned how, when, and why crustal faults either facilitated or inhibited magma transport to the surface. The main reason why magma became channelled along the faults is due to the mechanical properties and attitude of the faults. It was previously proposed that the crustal faults at Santorini are the main facilitators for magma to move from depth to the surface (Mountrakis et al. 1996). Through our various studies, we find this is a simplistic viewpoint and have provided a more complete analysis of which and how many of the mapped dyke segments (13%) used faults and the mechanisms under which they did this. We find that faults do not always facilitate the magma emplacement at Santorini; in fact, on many occasions, they do not, but that there are instead a complex set of variables that need to be met for the magma to use a fault zone.
The studied section of the northern caldera wall of Santorini was constructed by many different periods of effusive and explosive eruptive volcanic activity, which were generally fed by five radial subswarms of dykes. The sections of dykes which are exposed in the caldera wall are obviously very shallow, and hence they dip sub vertically since the eruption surfaces can, in many cases, also be observed and studied and since there has been virtually no vertical erosion of the top surface. If the dykes were fed by a much shallower magma chamber, then it is plausible that they would be inclined since the magma chamber would exert a stronger control on the stress field (Gautneb and Gudmundsson 1992). In that case, inclined sheets would form and potentially be even more less likely to use crustal faults in their paths towards the surface since the angle between the sheets and faults would be too great for the fault to capture the sheet (Browning and Gudmundsson 2015; Drymoni et al. 2021).

The crustal segment hosting the Santorini complex has experienced both faulting and dyke injection for the past several decades. It is the formation of dykes, their subsequent eruption and faulting related to magma chamber processes, and the local conditions of the stress field, rather than regional tectonics, which have shown to be the main mechanisms that control the current morphology of the island.

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