**Tectonics**

**RESEARCH ARTICLE**

10.1029/2020TC006641

**Key Points:**

- New and novel (re)interpretation of extension-related structures in Cycladic Blueschist Unit
- Extensional structures resulted from high degree of pure-shear flattening during general-shear deformation
- Structures are interpreted to reflect an extensional hinge zone in southern Cyclades

Supporting Information:

Supporting Information may be found in the online version of this article.

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Citation:

Ring, U., & Glodny, J. (2021). Geometry and kinematics of bivergent extension in the southern Cycladic archipelago: Constraining an extensional hinge zone on Sikinos Island, Aegean Sea, Greece. Tectonics, 40, e2020TC006641. https://doi.org/10.1029/2020TC006641

Received 23 NOV 2020
Accepted 20 MAY 2021

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**Geometry and Kinematics of Bivergent Extension in the Southern Cycladic Archipelago: Constraining an Extensional Hinge Zone on Sikinos Island, Aegean Sea, Greece**

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**Abstract**

We report the results of a field study on Sikinos Island in the Aegean extensional province of Greece and propose a hinge zone controlling incipient bivergent extension in the southern Cyclades. A first deformation event led to top-S thrusting of the Cycladic Blueschist Unit (CBU) onto the Cycladic basement in the Oligocene. The mean kinematic vorticity number ($W_m$) during this event is between 0.56 and 0.63 in the CBU, and 0.72 to 0.84 in the basement, indicating general-shear deformation with about equal components of pure and simple shear. The strain geometry was close to plane strain. Subsequent lower-greenschist-facies extensional shearing was also by general-shear deformation; however, the pure-shear component was distinctly greater ($W_m = 0.3–0.41$). The degree of subvertical pure-shear flattening increases structurally upward and explains alternating top-N and top-S shear senses over large parts of the island. Along with an increased coaxial deformation component, the strain geometry became oblate. Published quantitative data from nearby Ios Island are similar and both data sets define an extensional hinge zone between top-N extensional deformation across large parts of the central and northern Cyclades and top-S extensional deformation at the southern and western fringe of the archipelago. This extensional hinge zone is an important large-scale structure forming early in the history of lithospheric extension due to southward retreat of the Hellenic slab.

**Plain Language Summary**

The Aegean Sea in the eastern Mediterranean is a largely landlocked basin resulting from the southward retreat of the Hellenic subducting slab. The geometry of the extensional structures is debated. Our work on Sikinos Island and a comparison with structures from nearby Ios Island at the southern end of the Cycladic archipelago in the central Aegean extensional province suggests the existence of an extensional hinge zone of a bivergent extensional fault system. To the north of this symmetric hinge, the extensional faults are asymmetrically top-N displacing, whereas to its south the extensional faults are top-S displacing. The extensional hinge zone is an important large-scale structure in the central Aegean Sea and might have an equivalent in West Turkey. We favor a model in which the symmetric hinge formed early during lithospheric extension, moved southward in concert with the retreating Hellenic slab, and then monovergent top-N extensional faults formed in due course to the north of the extensional hinge.

1. **Introduction**

The Aegean Sea extensional province in Greece is an exceptional laboratory for studying extensional deformation above the southward retreating Hellenic subduction zone. Starting with the pioneering work of Lister et al. (1984), aspects of low-angle extensional shear zones and associated brittle detachments have been worked out in great detail (e.g., Fassoulas, 1999; Grasemann et al., 2012; Jolivet et al., 2010; Lee & Lister, 1992; Lister & Forster, 1996). Initially, these studies focused on the sense of shear and the timing of extension (Brichaus et al., 2010; Buick, 1991; Gessner et al., 2001; John & Howard, 1995; Soukis & Stöckl, 2013; Urai et al., 1990), potential displacements (Brichaus et al., 2006, 2007; Ring et al., 2001), the relation of extensional shearing with metamorphism, partial melting and magmatism (Beaudoin et al., 2015; Cao et al., 2013; Keay et al., 2001; Kruckenberg et al., 2011; Rabillard et al., 2017; Ring et al., 2018; Vanderhaeghe, 2004), and how various single extensional shear zones and faults (in general referred to as detachments...
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in the Cyclades) can be grouped into “detachment systems” (Grasemann et al., 2012; Jolivet et al., 2010; Ring et al., 2011).

At the scale of the archipelago, there is the North Cyclades detachment system at the northern fringe of the Cyclades (Jolivet et al., 2010) (Figure 1). This detachment system is well defined in the west, where different detachments operated between about 21 and 8 Ma (Brichau et al., 2008, 2007; Zeffren et al., 2005). Further east on Ikaria and westernmost Samos, the North Cyclades detachment system becomes younger with ages of about 15–7 Ma (Beaudoin et al., 2015; Kumerics et al., 2005) suggesting that it propagated to the east.

Figure 1. Aegean tectonic map and cross section with major extensional detachment systems, simplified cross section shows basin-and-horst structure of Aegean, which at Aegean-wide scale appears rather symmetric with asymmetric Cyclades horst; outcrop of Cycladic basement (red) and exposed migmatite (dashed red line) in central Cyclades also shown. TCT, Trans Cycladic Thrust; NCDS, North Cyclades detachment system; NPDS, Naxos-Paros detachment system; VD, Vari detachment; NID, North Ios detachment; S/WCDS, South/West Cyclades detachment system; CD, Cretan detachment; SDF, Simav detachment fault; CMCC, Central Menderes core complex; SMM, South Menderes monocline.
Jolivet et al. (2010) proposed that the North Cyclades detachment system can be correlated with the Simav detachment in West Turkey (Figure 1). However, the Simav detachment operated over a relatively short period of time between 24 and 19 Ma (Işık et al., 2004; Thomson & Ring, 2006) and thus at a significantly different time than the detachments in Ikaria and West Samos. Furthermore, Ring et al. (2017a) showed that there is a zone of relatively old (>15 Ma) zircon/apatite fission-track ages along the west Anatolian coastline interpreted to reflect a lack of Miocene detachment faulting there. The West Turkish detachments taper out to the west while the Aegean detachments have not yet laterally propagated into West Anatolia (Figure 1). In map view, this geometry defines two spatially disconnected extension provinces separated by the zone of old fission-track cooling ages (Figure 1). A simple three dimensional elastic model of two laterally tapering detachment fault systems should cause extension perpendicular to the slip directions of the two detachment systems (Bernhard Grasemann, written communication, 2011). Such Miocene E-W extension has in fact been reported from Samos Island in the easternmost Aegean by Gessner et al. (2011), Ring et al. (1999) and Roche et al. (2019). Therefore, we regard the North Cycladic detachment system to taper out in West Samos and not continuing into West Turkey.

To the south of the North Cyclades detachment system follow the mid/late Miocene Vari detachment (Ring et al., 2003; Soukis & Stöckli, 2013) and the early-to-late Miocene Naxos-Paros detachment system (Bargnesi et al., 2013; Brichau et al., 2006; Buick, 1991; Gautier & Brun, 1994; Lister et al., 1984; Urai et al., 1990; Vandenberg and Lister, 1996). All these detachment systems are top-N/NE displacing. At the southern and western end of the Cyclades, the geometry of the detachments is more complicated (see below for details). There is another top-N detachment in North Ios (Forster & Lister, 1999; Lister & Vandenberg, 1996), which may extend into Sikinos and Folegandros islands to the west (Augier et al., 2015; see below).

South and west of the top-N/NE-displacing Cycladic detachment systems, Ring et al. (2011) defined the South Cyclades detachment system, which encompasses detachments from South Ios into the western Cycladic islands (Figure 1). Grasemann et al. (2012) focused on the West Cyclades detachments and lumped them into the West Cycladic detachment system. Because Ring et al. and Grasemann et al. described the same detachment system but coined different terms for one and the same general structure, it might be useful to rename this detachment system as the South/West Cycladic detachment to avoid confusion.

There are different views as to when lithospheric extension in the Cyclades commenced. It is generally accepted that the onset of large-scale extension is linked to the retreat of the Hellenic slab (Jolivet & Brun, 2010; Ring et al., 2010). A proposition favored by Jolivet and coworkers claims that extension started by 30 Ma and was controlled by a slow-down of Africa Plate motion at about that time (Jolivet & Brun, 2010). To the best of our knowledge, an extensional detachment fault system at about 30 Ma has never been unambiguously documented and there are no supra-detachment sedimentary basins forming in the Oligocene across the Cyclades as a response to extension. The lack of Oligocene sedimentary basins would demand that upper-crustal extension lagged considerably behind ductile extension a depth. In contrast, Ring et al. (2010) claimed that extension commenced by 23–21 Ma. They base their inference on the geochronology of mylonites in extensional shear zones, low-temperature thermochronology constraining rapid cooling of the footwalls, and the oldest (Aquitanian) sediments in supra-detachment basins. Extensional deformation starting in the early Miocene has also been discussed in a West-Turkey context by Gessner et al. (2013).

### 1.1. Extensional Hinge Zone and Scope of This Study

The summary on the Cycladic detachment systems indicates that there needs to be a zone separating the top-N from the top-S detachments. We refer to this zone as an “extensional hinge zone.” Holohan (2007) described an “extensional hinge zone” associated with experimental caldera collapse. Recently, the term was used in connection with salt tectonics (McClay & Hammerstein, 2020). Calderas and diapirs are tectonic structures dominated by vertical movements and relatively small finite strains. Horst-and-graben structures in evolving rifts would also define an extensional hinge zone, that is, the axis of a horst bounded by sets of normal faults with opposite vergence (Jan Behrmann, written communication, 2021). We use a similar definition of an “extensional hinge zone”; however, in this article the “hinge” is demarcating the joint footwall (or structural dome) of a set of low-angle detachment faults with opposite vergence.
Two-dimensional numerical modeling work of large-scale extension of relatively weak, low-viscosity lower crust by Gessner et al. (2007) and Tirel et al. (2009) resulted in dome-shaped structures bounded by two opposite facing (bivergent) extensional shear zones. Distributed ductile flow of the lower crust is attracted into the extending region from both sides and is largely coaxial at the scale of the extension structure. The dome-shaped structure resembles an extensional hinge zone or culmination, called “apex” by Tirel et al. (2009). In the experiments, this geometry forms during early stages of extension while the lithosphere is relatively hot.

Tirel et al. (2009) compared their modeling results to the Cyclades and argued for three parallel detachment systems that developed as a result of Miocene extension. These three detachment systems are supposed to coincide with the North Cyclades, the Naxos-Paros, and the Ios-Folegandros detachment systems (Tirel et al., 2009). The first two systems are controlled by N-dipping detachment zones, while the deformation pattern on Ios suggests that the Ios-Folegandros detachment system has no marked asymmetry (Tirel et al., 2009).

The focus of this article is to (re)investigate the relationship between top-N/NE- and top-S/SW-displacing detachments in the southern Cyclades. An important question is how the two systems might have kinematically interacted with each other, that is, did the extensional structures result from two superimposed (near) simple-shear detachments or did they result from a high degree of pure-shear flattening during general-shear deformation as has recently been proposed for Ios Island by Mizera and Behrmann (2016). A study quantifying the degree of (non)coaxiality of deformation similar to that of Mizera and Behrmann (2016) on Sikinos Island may help to define an extensional hinge zone at the regional scale and discuss its tectonic significance.

2. Setting

The Hellenides are an arcuate orogen stretching from mainland Greece across the Aegean Sea region into western Turkey (Figure 1). They formed to the north of the Hellenic margin, along which NNE-ward subduction of the African plate beneath Eurasia is accommodated. In the Aegean Sea transect, the general architecture of the main tectonic units of the Hellenides is north dipping, the orogen youngs southward and can subdivided from top (north) to bottom (south) into: (a) The Rhodope-Balkan-Sarkaya Zone, (b) the Vardar-Izmir-Ankara Zone, (c) the Pelagonian-Lycian Zone, (d) the Cycladic Zone, and (e) the External Hellenides (including the Pindos, Tripolitza and Ionian zones) (Dürr et al., 1978; Figure 1).

The Rhodope-Balkan-Sarkaya Zone consists of continental fragments of the Eurasian plate, underneath which oceanic crust of Neotethys was subducted during Mesozoic convergence (Robertson et al., 1991). The related suture is the ophiolitic Vardar-Izmir-Ankara Zone. The underlying Pelagonian Zone comprises Paleozoic basement with a late Paleozoic and Mesozoic carbonate cover overlain by Jurassic ophiolite obducted toward the end of the Jurassic (Aubouin, 1959; Jacobshagen, 1986; Kilias et al., 2010).

To the south of the Pelagonian Zone is the Cycladic Zone. The dominant rock unit is the Cycladic Blueschist Unit (CBU) (Figure 2), which represents stretched continental fragments of the Adriatic Plate. The major tectonic members of the CBU are in structurally descending order: (a) A mélange-like assemblage of ophiolitic rocks and garnet-mica schist embedded in a serpentinicitic chlorite-talc schist matrix (Okrusch & Bröcker, 1990; Ring et al., 1999), (b) A Permo-Carboniferous to latest Cretaceous passive-margin sequence composed of marble, metapelite, quartzite, and metabasite that has been intruded by Triassic granitoids (Reischmann, 1997; Ring et al., 1999) (CBU passive-margin sequence). Below the CBU follows the early Paleozoic to Carboniferous Cycladic basement (Pouliaki et al., 2019; Reischmann, 1997), a nappe made up by orthogneiss, garnet-mica schist and dolomitic marble (Ring et al., 1999; Figure 2).

The various nappes of the CBU have undergone high-P metamorphism between about 53 and 30 Ma (see reviews in Jolivet & Brun, 2010, and Ring et al., 2010). The Cycladic basement experienced distinctly lower-grade high-P metamorphism (Franz et al., 1993; Gupta & Bickle, 2004). Because the CBU high-P rocks are tectonically on top of the lower-grade high-P rocks of the Cycladic basement, the contact must have been a thrust at some stage (Huet et al., 2009; Peillod et al., 2017; Peillod, Majka, et al., 2021). Both units were intruded by S- and I-type plutons between ~17 and ~11 Ma (Aitherr et al., 1982; Bolhar et al., 2010, 2012) during a regionally widespread greenschist-/ amphibolite-facies metamorphic overprint at ~22–11 Ma.
Miocene syn-extensional sediments occur above the CBU (Böger, 1983; Kuhlemann et al., 2004; Rösler, 1978; Sanchez-Gomez et al., 2002).

Further to the southwest in the External Hellenides, the weakly metamorphosed Pindos Zone and the Tripolitza Unit occur. The Cretan detachment tectonically separates the non-high-P Tripolitza Unit from the underlying, Miocene high-P rocks of the Ionian Unit (Fassoulas et al., 1994; Jolivet et al., 1996; Ring et al., 2001; Ring & Yngwe, 2018; Rosenbaum et al., 2007; Thomson et al., 1999). Above the Cycladic and Pindos zones are the lithologically heterogeneous upper units consisting of non-metamorphosed, Tertiary greenschist/blueschist-facies, and Cretaceous low-P rocks (Jacobshagen, 1986; Seidel et al., 1982).

In summary, the various tectonic units below the oceanic Vardar-Izmir-Ankara Zone in the Aegean transect of the Hellenide orogen show evidence for sustained high-P metamorphism. Reviews by Jolivet and Brun (2010) and Ring et al. (2010) showed that, in general, the age of this metamorphism becomes younger from north (east) to south (west) toward lower structural units and reflects stages of the southward propagation of the subducting Hellenic slab. Most of the exhumation of the high-P rocks occurred directly after their high-P metamorphism in extrusion wedges in the subduction zone (Chatzaras et al., 2006; Ring et al., 2007, 2020; Xypolias et al., 2003). However, the final stages of exhumation took place in the footwalls of the extensional detachments described above (Avigad et al., 1997; Beaudoin et al., 2015; Brichau et al., 2008, 2010; Ring et al., 2003).
3. Lithospheric Extension in the Southern Cyclades

The extensional hinge zone in the southern Cyclades is well characterized on Ios Island (Figure 1). Starting with the work of Vandenberg and Lister (1996), most workers concluded that the geometry of Miocene extensional deformation on Ios is bivergent, that is, top-N in the north and top-S in the southernmost part of Ios (Forster & Lister, 1999; Lister & Forster, 1996; Mizera & Behrmann, 2016; Thomson et al., 2009; Vandenberg & Lister, 1996). In South Ios, there are two sets of top-S structures, the first one developing under higher deformation temperatures (~450°C) than the second one, which proceeded under lower greenschist-facies conditions (<~350°C). The first event has been ascribed to thrusting of the CBU onto the Cycladic basement (Huet et al., 2009) and Rb-Sr multi-mineral age data by Thomson et al. (2009) and ^40Ar/^39Ar white-mica and potassium-feldspar dating by Forster and Lister (2009) and Forster et al. (2018) dated this event at 34.5 ± 2.5 Ma and 35–34 Ma, respectively.

The second set of lower-greenschist-facies top-S structures was dated by Thomson et al. (2009) at 19–18 Ma. Based on Rb-Sr dating of extensional mylonite along with low-T thermochronology, Thomson et al. (2009) further showed that top-N and top-S extensional deformation was proceeding in concert since at least 19–18 Ma and both footwalls cooled rapidly thereafter (see also Forster et al., 2018). To the best of our knowledge, the only kinematic study that did not report the top-S extensional detachment at the southern tip of Ios is Huet et al. (2009), who only described monovergent top-N extensional shear associated with back-arc extension. However, if top-N extension envisaged by Huet et al. (2009) is of the same age as the top-S structures as shown by the above studies, then both sets of extensional structures need to belong to the same tectonic event, rendering the Huet et al. (2009) proposition of monovergent top-N extension inconclusive. Furthermore, monovergent top-N extension is hard to reconcile with the quantitative data of Mizera and Behrmann (2016), who combined finite-strain and kinematic vorticity analysis demonstrating that ~70% N-S crustal stretching was accompanied by up to 40% of subvertical shortening. The very low kinematic vorticity number (0.26 on average) indicates almost pure-shear extensional deformation at the island scale. In other words, the extensional hinge zone is defined by a large degree of coaxial deformation expressed by alternating top-N and top-S shear-sense indicators (Mizera & Behrmann, 2016). Almost pure-shear flow in the basement and bivergent shear zones on Ios indicate that extension is dominantly coaxial, significantly reducing the tectonic overburden, but limiting the formation of discrete detachments. Nonetheless, Mizera and Behrmann (2016) argued that the noncoaxial component of shearing suggests that there has been relative movement on the order of 10–15 km between the Cycladic basement and the CBU in the hanging wall. Mizera and Behrmann (2016) associated this displacement with the top-S South Ios detachment and discussed that this top-S detachment formed part of a much larger system of Miocene-age detachments in the Aegean.

At the Cyclades scale, this hinge zone separates top-N detachments across most of the central and northern archipelago from the top-S detachments at the southwestern periphery of the Cyclades giving the bivergent extension system an overall asymmetric geometry (Figure 1).

This extensional hinge zone marks an important structure in the Cyclades. The close proximity of Sikinos and Ios prompted our interest in revisiting the extensional structures on Sikinos Island. It appears conceivable that Sikinos is also part of the South/West Cyclades hinge zone. Although Augier et al. (2015) showed the CBU/basement contact in South Sikinos to be a top-S ductile and brittle detachment (their Figure 1), they proposed a model of monovergent, simple-shear-type top-N extensional shearing on Sikinos (and also Folegandros) Island. Augier et al. (2015) reported top-S shear-sense indicators but did not consider them kinematically significant.

4. Geology of Sikinos Island

The geology of Sikinos encompasses the Cycladic basement tectonically overlain by the CBU passive-margin sequence (Augier et al., 2015; Avdis & Photiades, 1999; Franz et al., 1993; Gupta & Bickle, 2004; van der Maar, 1980; van der Maar & Jansen, 1983). The pre-Alpine basement consists of metagranodioritic gneiss and metaaplitic dikes, metavolcanics (quartz porphyry), garnet-mica schist and rare amphibolite (Andriessen et al., 1979; van der Maar, 1980). The magmatic rocks represent variably deformed Carboniferous (~325–310 Ma) calc-alkaline plutonic rocks emplaced into early Paleozoic paragneiss (Photiades &
Keay, 2003; Poulaki et al., 2019). The basement rocks are polymetamorphic with Carboniferous metamorphism, an early Tertiary high-P stage, followed by near-isothermal decompression and a greenschist-facies overprint (Augier et al., 2015; Franz et al., 1993; Gupta & Bickle, 2004). The CBU passive-margin sequence comprises calcitic and dolomitic marble and mica schist with intercalated metabasite (eclogite, blueschist, and greenschist) lenses. Augier et al. (2015) estimated the CBU sequence to be 6–7 km thick.

The contact between the Cycladic basement and the overlying CBU is a matter of debate. At both exposures of the basement, up to ~300 m of variably deformed granite and quartz porphyry is overlain by more intensely deformed paragneiss and quartz-mica schist (Augier et al., 2015; Avdis & Photiades, 1999; Franz et al., 1993; Gupta & Bickle, 2004). While the latter rocks were previously considered part of the pre-intrusive metasedimentary basement rocks, new detrital zircon U-Pb data revealed Permian depositional ages showing that they are part of the CBU (Poulaki et al., 2019). The detrital U-Pb zircon data suggest an uninterrupted progression in provenance signature across the CBU/basement contact suggesting erosional sourcing of the Permotriassic strata from underlying Carboniferous plutonic rocks. Poulaki et al. (2019) argued that the almost continuous “chronostratigraphic” succession supplies no evidence for an apparent structural break across the contact.

In contrast, the currently available P-T data suggest a distinct break in metamorphic P across the contact. Despite the fact that different methods have been used for obtaining P-T estimates, Franz et al. (1993) and Gupta and Bickle (2004) arrived at very similar P-T estimates for Alpine high-P metamorphism of the basement. Especially the P estimates show vastly different values for the basement (10–12 kbar) (Franz et al., 1993; Gupta & Bickle, 2004) and the overlying CBU (18–20 kbar) (Augier et al., 2015). A distinct metamorphic break at the Sikinos basement/CBU contact led Gupta and Bickle (2004) and Augier et al. (2015) to argue for significant displacement between basement and CBU.

5. Structural Data

In general, the metamorphic rocks in Sikinos are shallowlly N-dipping and dissected by N- and S-dipping normal, and NE-striking dextral oblique-slip faults (Figure 2). The late faults resulted from N-S extension (Figure 2; Augier et al., 2015). The northerly dip of the rocks suggests that Sikinos as a more or less continuous block was tilted north during late faulting. We studied the northern part of the CBU section, which is not dominated by marble. We also mapped the basement/CBU contact at Panteleimona Bay west of Alopionia in detail, and also at Agios Ioannis in South Sikinos.

5.1. Foliations, Stretching Lineations, and Folds in Cycladic Blueschist Unit

The main foliation, Sm, is primarily be made up by lower-greenschist-facies minerals (white mica, chlorite, albite and quartz in mica schist, chlorite, actinolite and albite in greenschist) and dips, on average, at 5°–30° to the N to NE (Figure 3) (Augier et al., 2015; Avdis & Photiades, 1999; Gupta & Bickle, 2004; van der Maar, 1980). This lower-greenschist-facies foliation overprints an earlier high-P foliation made up by glaucophane, omphacite, chloritoid, epidote, phengite, and quartz. Commonly, both foliations are subparallel to each other but in places one can observe that lower-greenschist-facies structures overprint an earlier foliation containing high-P relics.

On Sm, a N-S-trending stretching lineation, StrSm, marked by elongated quartz-albite aggregates, stretched pebbles in marble conglomerate, and the preferred alignment of mica and chlorite occurs (Figure 3). StrSm formed under lower-greenschist-facies metamorphic conditions. An earlier high-P stretching lineation, Strhp, marked by aligned glaucophane and phengite, as well as elongated chloritoid and omphacite is locally observed in blueschist, eclogite and quartzite. Strhp shows greater orientational scatter (Figure 3; Augier et al., 2015).

Relic, intrafolial, isoclinal folds with axes subparallel to the high-P stretching lineation are rare (Figure S1). Later, isoclinal folds deform the high-P foliation with Strhp and Sm is axial-planar to these FSm folds, which have N- to NE-trending axes (Figure 3). FSm depicts various geometries, common are isoclinal folds with shallowly to subhorizontal axial planes and rather straight fold axes parallel to StrSm. FSm sheath folds with subhorizontal axial planes are also observed. There can be an angle of up to 30° between FSm and StrSm, which
occurs in strongly deformed metapelitic lithologies where the angle between $F_M$ and $S_M$ may reflect local curvature of sheath-fold axes toward the fold hinges (Figure S1). Non-parallel attitudes between $F_M$ and $S_M$ also occur in areas of small strain intensity. $S_M$ is folded by gentle, large-scale upright $F_{M+1}$ folds with shallowly N/NW-plunging axes. Augier et al. (2015) suggested that these folds control the exposure of the Cycladic basement in cores of $F_{M+1}$ antiforms.

5.1.1. Shear-Sense Indicators in CBU

Kinematic indicators associated with the high-P foliation and $S_{HP}$ are rare. In outcrops of high-P rocks in North Sikinos near Agios Ilias, we observed top-S shear-sense indicators associated with a 170°-trending stretching lineation expressed by aligned and stretched glaucophane-epidote aggregates. At Agios Georgios Bay, blueschist has clasts composed of glaucophane, epidote, and carbonate and asymmetries of the clasts depict top-S shear (Figures 4a and 4b) but also alternating top-S and top-N shear senses (Figure 4c). Asymmetric quartz veins also provide top-S shear (Figure 4d). In some scattered outcrops of high-P rocks in Northeast Sikinos both top-S and top-N kinematic indicators were observed. All these shear-sense indicators developed when glaucophane, phengite, and epidote were stable (Figures 5a and 5b), probably during initial decompression of the high-P rocks.

By far most of the observed kinematic indicators developed under lower-greenschist-facies metamorphism as already stated by Augier et al. (2015), who found that lower-greenschist-facies extensional shearing was “extremely asymmetric as the Cycladic Blueschists unit is pervasively affected by top-to-the-north shearing deformation.” We also mapped top-N kinematic indicators in the CBU (which will, in part, be described be-
low in Section 5.2 as they relate to the CBU/basement contact). Most shear-sense indicators are shear bands and asymmetric clasts in mica schist (Figures 6a and 6b). In marble, we observed asymmetric dolomite boudins (Figures 6c and 6d). In general, lower-greenschist-facies top-N sense-of-shear indicators are more abundant in the deeper parts of the CBU above the basement contact.

Augier et al. (2015) also described top-S kinematic indicators on Sikinos Island and suggested that the top-S structures formed contemporaneous with top-N shear. The authors ascribed the coexistence of opposite kinematic indicators to a higher degree of coaxial deformation in less deformed domains and did not consider them kinematically important. Because of this, we focus here on top-S shear-sense indicators in some detail, as they represent the more controversial part of lower-greenschist-facies deformation on Sikinos.

In mica schist in central Sikinos, that is, between about Alopronia, Episkopi, Chora, Agios Ilias, and Agios Georgios (Figure 2), we mapped numerous top-S kinematic indicators (Figure 7). These include asymmetric carbonaceous, mafic clasts in mica schist (Figures 7a and S2a) and abundant shear-band structures (Figures 7b–7d), the latter especially about 1 km west of Agios Ilias. Here, the shear-band structures show a progressive evolution from lower-greenschist-facies to semi-ductile and cataclastic deformation behavior (Figure 7d). In marble, asymmetric clasts of various size occur (Figure S2b). We observed the asymmetric phacoidal structures in marble especially south of Episkopi (Figure 3).

Augier et al. (2015, their Figure 2) mapped the top-N Voudhia shear zone in Northeast Sikinos (Figure 3). There is no deflection of $S_m$ into the Voudhia shear zone and $S_m$ and $S_{tr_M}$ are well expressed in all schist outcrops and some marble layers, especially in marble conglomerate, outside the shear zone. While we agree that top-N shear bands occur in and near the Voudhia shear zone, we also mapped numerous top-S

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**Figure 4.** Kinematic indicators associated with high-P metamorphism and initial decompression in CBU; all photos from Agios Georgios Bay (36°41′36″N, 25°10′04″E). (a) Asymmetric clasts composed of glaucophane, epidote, omphacite and ankeritic carbonate with strain shadows indicating top-S shear. (b) Similar but distinctly more epidote-rich clasts as (a) also showing top-S shear. (c) Epidote-rich blueschist with abundant blueschist and quartzite clasts; asymmetric strain shadows around clasts supply both top-S and top-N shear sense. (d) Asymmetric quartz vein in epidote-rich blueschist similar to (c) providing top-S shear sense. CBU, Cycladic Blueschist Unit.
Figure 5. Thin-section photographs. (a), (b) High-P rock sample SIK20-11 from Agios Georgios Bay. (a) Asymmetric, quartz-filled strain shadow around glaucophane; PPL. (b) Phengite-filled shear band between glaucophane; PPL. (c) Potassium feldspar with simple twinning with asymmetric, top-S strain shadows (arrows) filled with white mica and quartz; incipient recrystallization of feldspar (red arrow); sample SIK20-16 (36°39′42″N, 25°07′16″E), XPL. (d) Potassium feldspar synkinematically disintegrating into white mica (white arrows); associated kinematic indicators yield top-S shear sense; sample SIK20-6 (36°39′36″N, 25°07′18″E), XPL. SIK20-16 and SIK20-6 from Panteleimona Bay. PPL, plane-polarized light; XPL, cross-polarized light; gl, glaucophane; phe, phengite; wm, white mica; qtz, quartz; fsp, feldspar.

Figure 6. Greenschist-facies top-N kinematic indicators associated with Sm in CBU. (a) Series of N-dipping shear bands in carbonaceous mica schist W of Alopromia (36°40′23″N, 25°08′19″E). (b) Same set of N-dipping structures as in (a) (36°40′24″N, 25°08′22″E). (c) Boudinage of dolomite layer in calcite marble showing asymmetric top-N geometry (36°39′54″N, 25°07′35″E). (d) Same general outcrop as in (c) showing asymmetric dolomite clast indicative of top-N shear.
shear indicators (Figures 7e and 7f, see also Figures S2c–S2f). Most of the unambiguous kinematic indicators are asymmetric quartz veins (Figures 7e, 7f, S2c, and S2d). Shear bands and low-angle normal faults (Figure S2f) occur as well.

Figure 7. Greenschist-facies top-S kinematic indicators associated with S\textsubscript{m} in CBU. (a) Asymmetric ankeritic mafic clast in carbonaceous schist at Alopronia port (36°40′29″N, 25°08′34″E); clast similar to those in Figure 4a but mineralogy includes epidote and chlorite, not glaucophane and omphacite. (b)–(d) Top-S shear bands in carbonaceous chlorite schist at dirt road to Agios Ilias (36°42′15″N, 25°08′11″E); shear bands formed at lower-greenschist-facies conditions and cataclastically overprinted in (d), gray shading indicates cataclastic layer. (e), (f) Greenschist-facies top-S kinematic indicators in Voudhia shear zone; both outcrops within 200 m of 36°42′26″N, 25°09′46″E. (e) Asymmetric quartz boudin in chlorite-mica schist. (f) Asymmetric quartz boudin in quartz-albite schist (see also red arrow in Figure S2c). CBU, Cycladic Blueschist Unit.
5.2. Cycladic Basement and its Contact With the CBU

The exposure of the Cycladic basement west of Alopronia (Figure 2) is largely controlled by a series of late N-dipping normal faults (Figures 8, 9a, and 9b). The actual contact is a semi-ductile to cataclastic shear zone (Figure 9c). Augier et al. (2015) noted that the orientation of Sm and Sm' remain unchanged across this contact.

Kinematic indicators at the semi-ductile to cataclastic contact are rare and often hard to interpret. Shear bands and asymmetric strain shadows around small feldspar and quartz clasts indicate top-N shear, although there are outcrops with alternating top-S and top-N kinematic indicators (Figure 9d). Riedel structures in foliated cataclasite show top-N shear. A zone of lower-greenschist-facies mylonite occurs in mica-chlorite-albite schist about 200–300 m north of the basement/CBU contact (Figure 8), where top-N shear-sense indicators are ubiquitous (Figures 9e and 9f). In this mylonite belt, we mapped 37 kinematic indicators, 32 of which provided top-N shear.

Structurally about 100–150 m below the basement/CBU contact, top-S kinematic indicators become more abundant, as was already noted by Augier et al. (2015). However, the basement/CBU contact northwest of Panteleimona Bay is a mylonitic top-S shear zone (Gupta & Bickle, 2004). The top-S kinematic indicators (Figure 10) developed at greenschist-facies conditions and most are associated with garnet breaking down to chlorite. In places, there is incipient recrystallization of potassium feldspar in some top-S shear fabrics (Figure 5c). Top-N shear-sense indicators also overprint top-S ones in quartz porphyry at the northern and western ends of the orthogneiss outcrops at Panteleimona Bay. The top-N kinematic indicators are of lower metamorphic grade than the top-S ones (Figure 5d).

We envisage that the preservation of the top-S shear zone at the basement/CBU contact (Figure 8) is due to its N/NNW strike, which was unsuitably oriented for N-S shearing during subsequent lower greenschist-facies and cataclastic reworking.

6. Finite Strain and Kinematic Vorticity

Especially for the lower-greenschist-facies deformation event there are outcrops in which alternating top-S and top-N shear-sense indicators occur next to each other (Figures 9d and 11a) and there are outcrops with symmetric structures (Figures 11b and 11c). These examples suggest at least localized areas where tectonic flow under lower-greenschist-facies conditions was coaxial or close to coaxial. However, alternating top-N and top-S shear senses occur throughout the CBU section over large parts of Sikinos Island.
suggesting that coaxial flow is not a local phenomenon. There are also indications for deviations from plane-strain deformation. Chocolate-tablet boudinage is observed, especially in marble. The structure shown in Figure 11d shows extensional fibers in various directions and mutual cross-cutting relations. This suggests extension in two principal strain directions and thus oblate-strain geometry. In outcrop, S ≈ L fabrics (which means the intensity of foliation (S) development is distinctly more pronounced than the development of the stretching lineation, StrM (L), in this case) are common, although S ≈ L fabrics and L > S

**Figure 9.** (a) Annotated photo of basement/CBU contact W of Alipronia; drag of layering in marble toward N-dipping normal fault. (b) Normal fault between orthogneiss (left) and garnet-mica schist (right) in basement. (c) Cataclastic contact between marble and fine-grained basement orthogneiss; note that foliation remains subparallel to each other across contact (36°39′32″N, 25°07′13″E). (d) Augengneiss with potassium-feldspar porphyroclasts, asymmetries of clasts supply alternating top-N and top-S shear sense, and symmetric structures; some clast rotation interferes with each other (36°39′39″N, 25°07′17″E). (e), (f) Top-N shear bands in chlorite-mica-albite schist (36°39′52″N, 25°07′16″E). CBU, Cycladic Blueschist Unit.
fabrics also occur. Finally, as already noted by Augier et al. (2015), the main $S_M$ foliation is hardly deflected at the basement/CBU contact and in the Voudhia shear zone, suggesting a distinct component of coaxial flattening perpendicular to the shear-zone boundaries.

For testing the significance of (near) coaxial flow and non-plane-strain deformation, we quantified finite strain and the mean kinematic vorticity number ($W_m$). Finite strain was estimated by the $Rf/\phi$ and Fry methods (Dunnet, 1969; Ramsay & Huber, 1983); $W_m$ by the porphyroclast aspect ratio (PAR) method (Bailey & Eyster, 2003; Grasemann et al., 1999; see review in Xypolias, 2010). The methods have been summarized in Kumerics et al. (2005) and Ring and Kumerics (2008), and are provided in the data supplement (with Figure S3 showing examples of deformed conglomerate). As stated in Xypolias (2009, 2010) and Law (2010), the PAR method records the entire ductile deformation history since the porphyroclasts add rotational increments during the deformation (e.g., Passchier, 1987, 1988) and should provide robust results close to the real mean vorticity of flow. Furthermore, Kumerics et al. (2005) reported kinematic vorticity from the footwall shear zone of a segment of the North Cycladic detachment system on Ikaria Island (Figure 1) using the PAR method, which are consistent with $W_m$ values obtained by other methods in the same rocks (see discussion of its significance in Xypolias, 2010), providing further support for the robustness of the PAR method. Pure and simple-shear components are equal at $W_m = 0.71$ (Law et al., 2004; Means et al., 1980).

In general, the finite-strain data show oblate-strain geometry, with a few data points in the prolate field (Figure 12a). This result in is line with the $S > L$ fabrics and other structures reported above. Fry analyses from quartz porphyry and augen granite with top-S shear-sense indicators supply data close to the plane-strain line. Strain intensity, as displayed by the distance of the data points from the origin, appears slightly higher in the oblate samples.
The kinematic vorticity data provide low $W_m$ of 0.3–0.41 for CBU rocks dominated by $S_M$/Str$_M$ structures. CBU rocks showing hardly any greenschist-facies overprint and early top-S structures provide distinctly higher $W_m$ of 0.56–0.63 (Figure 12b), which still indicate a pure-shear dominated flow regime. The top-N structures near the top of the basement have $W_m$ 0.58–0.73, but some data from outcrops with coaxial fabrics (e.g., Figure 8d) provide considerably lower $W_m$ (0.16–0.23) indicating virtually pure-shear deformation (Figure 12c). Basement outcrops with early top-S fabrics show highest $W_m$ values of 0.72–0.84.

The kinematic vorticity gets larger toward structurally lower levels for the lower-greenschist-facies $S_M$/Str$_M$-dominated structures, that is, pure-shear deformation becomes less toward the CBU/basement contact, the latter of which is characterized by about equal pure- and simple-shear components. The earlier high-P fabrics depict higher $W_m$ between 0.56 (CBU) and 0.84 (basement) indicating a greater simple-shear contribution to general shearing.

Augier et al. (2015) emphasized that Carboniferous granodiorite dikes are subparallel with $S_M$ at the CBU/basement contact, but the dikes can also be subperpendicular to $S_M$ (Augier et al., 2015, their Figure 3c). The rotation and thinning of the dikes indicate very high finite and shear strains (cf., Norris & Cooper, 2003) and the orientation of the dikes into parallelism with $S_M$ is strongly aided by the pure-shear component of deformation (cf., Pfiffner & Ramsay, 1982). It is unknown whether or not the high strains are due to top-S shearing or top-N reactivation or a combination of the two.

There are not many quantitative studies from the Cyclades to which to compare the results. Mizera and Behrmann (2016) reported near plane strain associated with almost pure-shear flow on Ios. Kumeries et al. (2005) described generally oblate-strain geometry associated with $W_m$ of 0.13–0.8 from a segment of the North Cycladic detachment system on Ikaria. However, the extensional shear zone on Ikaria has consistent top-NNE kinematic indicators. Ring and Kumeries (2008) reported 75 finite-strain results from
across the Cyclades with 47 data points and the tensor average of all data plotting in the oblate field. Peilod, Tehler, et al. (2021) also described oblate-strain geometry from the CBU on Naxos and suggested that extensional shearing had a distinct pure-shear component. Xypolias (2010) reported near-plane-strain geometry from the lowermost CBU (and $W_m$ of 0.8–0.9) and prolate-strain geometry from the top of the CBU section ($W_m$ of 0.6–0.9) on Evia Island. However, Xypolias et al. claimed that their data relate to thrusting.

7. Tectonic Interpretation

The CBU/basement contact is an important tectonic structure on Sikinos Island. Recently some controversy arose because Poulaki et al. (2019) suggested that it represents an original "(pseudo)stratigraphic" boundary. However, the fact that basal CBU section on Sikinos was sourced from Cycladic basement does not necessarily indicate that the CBU section was sourced from the now directly underlying Sikinos basement. Outcrops of Cycladic basement are widespread across the central and southern Cyclades (e.g., Ios, Paros, Naxos) (Figure 1) and probably also occur below the Aegean Sea in between these islands. Any of this basement could have sourced the Sikinos basal CBU section rendering the Poulaki et al. "(pseudo)stratigraphic" boundary proposition uncertain. While a "(pseudo)stratigraphic" contact could potentially be in line with a large amount of vertical flattening and near-coaxial deformation, we believe that the pronounced P-T break across the contact is not a methodological artifact and in line with a tectonic nature of the contact.

7.1. Top-S Deformation

The high-P stretching lineations show some scatter (Figure 3), which is due to isoclinal folding of $\text{Str}_{HP}$ by greenschist-facies $F_M$ folds about axes that may have angles of up to 30° to $\text{Str}_{HP}$. Isoclinal $F_M$ folds with axes subparallel to $\text{Str}_{HP}$ cause an inversion of the shear sense in the "inverted" limbs of the folds. We envisage that this process caused the occasional top-N high-P shear-sense indicators in the CBU in Northeast Sikinos. However, post-$\text{Str}_{HP}$ isoclinal folding cannot explain the alternating top-S and top-N kinematic indicators at Agios Georgios Bay. The latter are better explained by the pure-shear component of deformation (Figure 12b).

Early top-S kinematic indicators also occur in the Cycladic basement (Augier et al., 2015; Gupta & Bickle, 2004). Top-S deformation is close to plane strain and has a distinct simple-shear component (Figure 12c). We propose that top-S shearing in the CBU and the basement are tectonically linked and resulted from top-S emplacement of the CBU during decompression onto the Cycladic basement. Given that this event put 18–20 kbar rocks on top of 10–12 kbar basement rocks suggests a thrust contact and that this thrust moved while the CBU high-P rocks decompressed by 6–10 kbar to explain P of 10–12 kbar in the basement. Evidence for top-S thrusting of the CBU onto the basement was reported by Huet et al. (2009) from nearby Ios Island and Peilod et al. (2017) from Naxos Island. The available age data from both islands show that thrusting occurred between about 35 and 30 Ma (Forster et al., 2018; Peilod et al., 2017; Thomson et al., 2009).
7.2. Lower-Greenschist-Facies Deformation

Our data show that the lower-greenschist-facies $S_M$ event is characterized by alternating top-N and top-S kinematic indicators. In general, this finding is in line with Augier et al. (2015), who ascribed the coexistence of opposite kinematic indicators to a higher degree of coaxial deformation in less deformed domains. Augier et al. (2015) considered top-N shear bands as representing the more ubiquitous simple-shear criteria.

Quantitative kinematic vorticity data aid an informed interpretation of the significance of the alternating shear senses (Bailey and Eyster, 2003; Law et al., 2013; Mizera & Behrmann, 2016; Ring et al., 2015; Xypolias, 2009, 2010). Especially in the upper CBU in North Sikinos, the kinematic vorticity data in outcrops with $R_{zz}$ ratios of 3–4.5 still record distinctly coaxial deformation ($W_m$ between 0.3 and 0.41) (Figure 12b). There is no evidence that deformation in North Sikinos is anywhere close to simple shear. It follows, that the alternating kinematic indicators indeed reflect a high degree of coaxial deformation. This coaxial deformation is characteristic for at least the northern half of Sikinos Island and therefore alternating shear senses are also representative for the upper parts of the CBU section. The gentle, large-wavelength $F_{M+1}$ did not affect the $S_M$ shear sense as they do not overturn the tectonostratigraphy.

Toward structurally lower levels approaching the CBU/basement contact in central and southern Sikinos, the pure-shear component of deformation becomes less. The actual CBU/basement contact is characterized by about equal pure- and simple-shear components ($W_m$ up to 0.73), which still explains lower-greenschist-facies top-S sense-of-shear criteria close to the contact. The only zone of intense shear we mapped that has almost monovergent top-N kinematic indicators is situated $\sim 200–300$ m above the CBU/basement contact, but no vorticity data exist for this zone.

The lower-greenschist-facies general-shear deformation involved a considerable degree of subvertical flattening resulting, in general, in oblate strain geometry. Near the CBU/basement contact, the simple-shear component is more pronounced and general-shear deformation had overall top-N tectonic transport. Above this contact zone, the pure-shear component dominated resulting in alternating shear senses.

It is generally agreed on that the lower-greenschist-facies deformation resulted from lithospheric extension due to retreat of the Hellenic slab (Augier et al., 2015; Grasemann et al., 2012; Jolivet et al., 2010; Ring et al., 2011). The age of this deformation event is unknown on Sikinos but was underway by at least 19–18 and lasted until 9 Ma on nearby Ios Island (Thomson et al., 2009).

8. Extensional Hinge Model

Shear-sense, quantitative finite-strain and kinematic-vorticity data for lower-greenschist-facies deformation on Sikinos are similar to the same set of quantitative data from nearby Ios Island (Mizera & Behrmann, 2016). Both data sets indicate a large component of coaxial stretching during Miocene extensional deformation. Furthermore, both data sets emphasize the importance of quantitative data for analyzing and interpreting heterogeneous and complex structural data with conflicting kinematic indicators.

On the island of Ios, an extensional hinge zone between a S-dipping shear zone in the south and a N-dipping shear zone in the north of the island is well defined. The culmination of Cycladic basement is the footwall to both detachments of the bivergent system (Figure 13a). Because the final stages of ductile extensional shearing at the top-S and the top-N extensional detachments are contemporaneous at 19–18 Ma (Thomson et al., 2009), the domal outcrop of the Ios basement describes the extensional hinge.

The above summarized numerical models by Gessner et al. (2007) and Tirel et al. (2009) support the conclusion that both detachments of the bivergent extensional system developed simultaneously at the onset of extension. Based on field studies in the Basin-and-Range extensional province in the western USA, Reynolds and Lister (1990) suggested that a bivergent system may be due to the main mylonitic zone rolling over to a dip opposed to the direction of upper-plate transport and an antithetic shear zone developing during the final stages of mylonitization on the opposite side. We do not consider such a scenario applicable to the Ios/Sikinos hinge zone because the available isotopic data are not compatible with a late-stage formation of one of the extensional detachments.
Figure 13. Tectonic interpretation. (a) Ios/Sikinos extensional hinge model illustrating coaxial deformation between bivergent top-S and top-N extensional fault system; geometry well-defined on Ios Island, on Sikinos coaxial deformation in CBU section was translated onto basement during top-N extension at general-shear CBU/basement contact (inset). (b) Bathymetric map of Cyclades and Menderes Massif of western Turkey showing major extensional detachments (red; same abbreviations as Figure 1) cut by high-angle normal faults (black) according to Sakellariou et al. 2016 (map from Aster and Emodnet data); also shown extensional hinge zones (dashed purple lines) discussed in text. Note that normal faults between Naxos and Mykonos, and to N of Sikinos extend eastward delimiting Sikinos-Ios-Iraklia (I)-Koufonisi (K)-Amorgos (SIIKA) horst. Dashed N-S line indicates cross section in (c). (c) Cyclades-wide N-S cross section with main extensional detachments and bivergent hinge zone at southern end of Cyclades (Ios, Sikinos); geometry of various detachments making up North Cyclades detachment (Jolivet et al., 2010) according to Brichau et al. (2007); note general N-verging asymmetry at Cyclades scale; cross section modified from Augier et al. (2015). (d and e) Comparison of Aegean Sea hinge and Menderes Massif hinge development since Oligocene/Miocene boundary; initial situation somewhat similar but extension in Menderes Massif localized at single Simav detachment; major differences in Pliocene when symmetric Central Menderes core complex formed while Aegean Sea region underwent block faulting. Abbreviations as Figure 1.
The data by Mizera and Behrmann (2016) showing plane-strain (near) coaxial deformation are mainly from the Cycladic basement. For the Sikinos basement, our data indicate the same strain regime on both islands. However, most of our data are from the tectonically overlying CBU. If the noncoaxial component of general-shear deformation caused top-N displacement at the basement/CBU contact, then most of the pure-shear-dominated deformation of the overlying CBU was acquired further south (Figure 13a, inset). This would project the Sikinos CBU section above the Ios basement for which Mizera and Behrmann (2016) quantified ~70% N-S stretching and up to 40% subvertical shortening during extensional flow (Figure 13a). This makes it likely that the Sikinos CBU/basement section represents the central and northern edge of the “Ios extensional hinge zone”.

This extensional hinge zone marks an important structure in the Cyclades and the question arises as to the lateral extent of this zone. For answering this question quantitative structural field studies are needed, especially from Folegandros Island to the west and the islands of Iraklia, Keros, Antikeri, and Amorgos to the east of Ios/Sikinos (Figure 13b).

Ages for extensional shearing suggest that top-S extension across most of the South/West Cyclades detachment system was underway by 21–19 Ma and proceeded until about 8–6 Ma (Coleman et al., 2020; Grasemann et al., 2012; Igléseder et al., 2011; Schneider et al., 2018; Thomson et al., 2009). Only for Serifos and Sifnos, ages appear to be younger at 13–6 Ma (Brichau et al., 2010; Ring et al., 2011). These ages are largely similar to ages for the inception of the North Cyclades, the Naxos-Paros, the N-Ios and related detachment systems (Bargnesi et al., 2013; Brichau et al., 2007, 2006; Laurent et al., 2017; Ring et al., 2009; Seward et al., 2009; Zeffren et al., 2005). Again, extensional shearing and faulting on some islands is younger (e.g., Ikaria, 15–8 Ma Beaudoin et al., 2015; Kumerics et al., 2005, Syros, 15–9 Ring et al., 2003; Soukis & Stöckli, 2013, Mykonos, 13–9 Ma). Most of the isotopic ages reflect the waning stage of shearing, not necessarily the onset of it. This brief summary shows that the extensional hinge is an early feature forming at the start of lithospheric extension in the Cyclades.

It is unclear what guided the development of the proposed extensional hinge zone. The kinematic-vorticity data (Mizera & Behrmann, 2016; this study) show a high degree of coaxial vertical flattening and alternating shear senses. However, the data by Kumerics et al. (2005) from Ikaria Island also show low kinematic vorticity but the shear sense during extensional shearing at the Ikaria segment of the North Cyclades detachment system is predominantly top-N (see also Beaudoin et al., 2015). A similar, but less well constrained case can be made for the Naxos-Paros detachment system for which Peillod, Tehler, et al. (2021) showed a distinct degree of coaxial deformation during extensional shearing associated with a dominant top-N sense of shear. Collectively, these data support distinct vertical flattening during large-scale ductile extension but pronounced vertical flattening does not appear a limiting factor causing the extensional hinge zone to form.

The onset of extension, and the extensional hinge, preceded the development of arc magmatism. It also preceded decompressional melting and migmatization as the decompression would be a result of extension. It is therefore tempting to relate the development of the extensional hinge to initial subduction and crustal thickening by thrusting. Grasemann et al. (2018) proposed the large-scale Trans Cycladic Thrust in the southwestern Cyclades (Figure 1) putting high-P rocks ≥18 kbar on top of high-P rocks characterized by ∼9–12 kbar. This situation is very similar to the basement/CBU contact on Sikinos. The same relationship might be true for the Ios basement/CBU contact but P-T conditions there are less well constrained. If so, the Trans Cycladic Thrust would be domed-up (or extensionally hinged) on Ios. It appears conceivable that the Trans Cyclades Thrust provided the structural heterogeneity that guided the subsequent development of the extensional hinge zone.

A chicken-and-egg question would be whether the updoming of the Ios basement was a result of bivergent extension or was actually critical for the development of the hinge zone in the first place? The Ios, and also the Sikinos, basement, was undergoing greenschist-facies metamorphism in the Miocene (Grütter, 1993) and is the structurally deepest exhumed unit in the southern Cyclades. The same basement is migmatitic further north on Naxos and Paros. It is conceivable that the top-N detachments on Ios and Naxos might be one and the same detachment and the deeper levels of the footwall are exposed in Naxos and Paros.

Despite the early development of the envisaged extensional hinge zone and symmetric extension at the Ios/Sikinos scale, lithospheric extension at the scale of the Cyclades is asymmetric (Augier et al., 2015; Jolivet et al., 2016; Grasemann et al., 2012; Igléseder et al., 2011; Schneider et al., 2018; Thomson et al., 2009).
& Brun, 2010) (Figure 13c). It appears plausible that the extensional hinge formed early and then migrated southward with the retreat of the Hellenic subduction zone, which experienced a sudden and rapid phase of rollback at about 23–21 Ma (Ring et al., 2010). We speculate that the bivergent detachment system with the extensional hinge developed first and then migrated southward in the direction of slab retreat. Subsequently, the Naxos-Paros and North Cyclades detachments formed in due course within 1–2 Myr (Figure 13d). We admit that such a proposition is hard to prove given the currently available age data.

There are distinct differences in extensional deformation between the Aegean Sea region and the Menderes Massif in West Turkey (Gessner et al., 2013; Ring et al., 1999). However, in both regions extensional deformation commenced at about the same time at the Oligocene/Miocene boundary. Furthermore, across the entire Menderes Massif, the overall extension geometry is also asymmetric with distinctly more important top-N shearing and faulting (Gessner et al., 2013; Isik et al., 2004; Thomson & Ring, 2006). Ring et al. (2017b) documented the early Miocene South Menderes Monoclone at the southern end of the massif, which formed by differential exhumation and uplift possibly due to unloading of the large-displacement top-N Simav detachment footwall. If so, the Simav detachment/South Menderes Monoclone would also represent an early Miocene extensional hinge (Figure 13e). An important difference to the Aegean Sea region is that in the Menderes Massif there is another, Pliocene extensional hinge since about 5 Ma (Hetzel et al., 1995; Gessner et al., 2001, 2013), which geometrically is akin to the symmetric extension in the Ios/Sikinos hinge zone.

9. Concluding Remarks

A field study on shear-sense indicators, quantitative finite-strain and kinematic-vorticity data reveal that Miocene extensional shearing on Sikinos Island was by general-shear deformation. The degree of subvertical pure-shear flattening increases structurally upward explaining alternating top-N and top-S shear senses over considerable parts of the island. Published quantitative data from nearby Ios Island are similar and define an extensional hinge zone between top-N extensional deformation across large parts of the central and northern Cyclades and top-S extensional deformation at the southern and western fringe of the archipelago. Extensional shearing was preceded by Oligocene top-S thrusting of the CBU onto the Cycladic basement and this thrust geometry may have provided the structural heterogeneity for the development of the extensional hinge zone.

Data Availability Statement

Supplementary data are available at https://doi.org/10.5281/zenodo.4877095, data comply with FAIR Data guidelines.

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