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ABSTRACT

Model studies of continental collision tectonics and pressure-temperature-time (*P-T-t*) predictions quantitatively demonstrate the key role of synsubduction uplift in the preservation of blueschist- and eclogite-facies metamorphic rocks. Rheological stratification of the subducting lithosphere allows development of detachment faults at the compositional boundary between the upper and lower crust. Material of supracrustal continental affinity is subducted to depths in excess of 50 km and then delaminated from the downgoing slab. Subduction continues while the detached domain is uplifted independently from the newly active hanging walls and footwalls. Two-dimensional thermal calculations indicate that the nearby active subduction provides the cooling mechanism required to preserve blueschists and eclogites, whereas the differential motions of the independently uplifted domains can explain the observed heterogeneity in *P-T-t* pathways, with some domains displaying a greenschist overprint.

INTRODUCTION

When subducted to depths >30 km, supracrustal rocks recrystallize at blueschist and eclogite metamorphic grade, forming an important element in the process of continent-continent collision (e.g., Ernst, 1975; Molnar and Gray, 1979; Hsü, 1991). During exhumation, such rocks will be significantly overprinted because of radioactive heat production unless there is a mechanism that provides sufficient cooling. Current models invoke the destruction of overlying crust by extension triggered by the buoyant rise of continental material from below (Rubie, 1984; Platt, 1986) or by simple shear driven by convection in the upper mantle (Wernicke, 1981; Davis, 1983). Cloos and Shreve (1988) proposed a subduction-channel model permitting the return of fluids and sediment from deep levels along the base of the accretionary wedge.

We present a model in which delamination of supracrustal rock layers during subduction occurs along rheologically weak zones in continental lithosphere. Subduction of the denser lower crust and mantle lithosphere continues while uplift of the lighter supracrustal slab occurs by reversal of movement along the subduction-zone fault system. Exhumation of very deeply buried rocks along such faults does not require complete destruction of initially overlying continental crust and mantle.

The model is applied to the Aegean Sea (Fig. 1) where Neogene extension (McKenzie, 1978; Le Pichon and Angelier, 1979; Angelier et al., 1982; Lister et al., 1984; Buick, 1991) and the *P-T-t* history of Eocene-Miocene high-pressure (high-*P*) metamorphism of rocks of the Attic Cycladic blueschist belt and Crete (Altherr et al., 1979; Seidel et al., 1982; Schliestedt et al., 1987; Wijbrans et al., 1990) are well documented. Seismic tomography has imaged the subducted slab to some 600 km depth, suggesting continuous subduction at the Hellenic trench (indicated by H in Fig. 1) from at least the Oligocene (Spakman et al., 1988). The presence of high-*P* rocks of early Tertiary age near the trench led Wijbrans and McDougall (1988) to suggest that subduction occurred throughout most of the Tertiary. In this paper the available *P-T-t* data from Sifnos (Schliestedt et al., 1987; Wijbrans et al., 1990) are used to define a two-dimensional numerical model of the tectonic evolution of Sifnos during the Tertiary.

PETROLOGY AND GEOCHRONOLOGY

Sifnos pressure-temperature-time data reflect the tectonics of Alpine accretion and subsequent extension, providing parameters for modeling the evolution of the Attic Cycladic belt as a whole. There are two different structural-metamorphic domains on Sifnos (Schliestedt et al., 1987; Fig. 2). In the north (domain I), blueschist high-*P*–low-*T* conditions were reached prior to 41 ±5 Ma, when these rocks cooled to ~360 °C (our estimate for the closure temperature for K-Ar in phengite). The rocks exposed in central Sifnos (domain II) contain mica relicts indicating a similar high-*P*–low-*T* metamorphic event, but showing markedly younger phengite Ar-closure ages of 30 ±2 Ma (Wijbrans et al., 1990). Domain II has been overprinted by greenschist minerals, for which temperatures locally in excess of about 450 °C have been estimated (Schliestedt et al., 1987), including white micas that crystallized (or cooled to ~360 °C) at 18.9 ±0.3 Ma, as determined by $^{40}Ar/^{39}Ar$ age-spectrum analysis (Wijbrans et al., 1990). At 7 ±2 Ma, the rocks of the Cycladic blueschist belt reached the surface (Wijbrans and McDougall, 1988). Although fluid infiltration occurred during Miocene greenschist overprinting in central Sifnos (Matthews and Schliestedt, 1984; Ganor et al., 1989), the Ar age spectra suggest that heat advection by fluid infiltration was local.

The petrologic data (Schliestedt et al., 1987) are summarized in Figure 3. Box A indicates the field of *P-T* conditions necessary to produce the high-*P*–low-*T* metamorphism observed in domain I and inferred for domain II. Box B is the field inferred for conditions during greenschist overprinting. Because maximum temperatures for both events were very similar, the transition from blueschist to greenschist metamorphism was originally described as isothermal decompression (Matthews and Schliestedt, 1984). Wijbrans et al. (1990) argued that isothermal decompression cannot be easily reconciled with the range in phengite K-Ar ages discussed above.

NUMERICAL MODEL AND RESULTS

To model *P-T-t* paths defined by the isotope and metamorphic data, we use a 900 and 300 km finite-difference mesh with a grid spacing of 1.5 km. Tectonic domains within the mesh are delineated

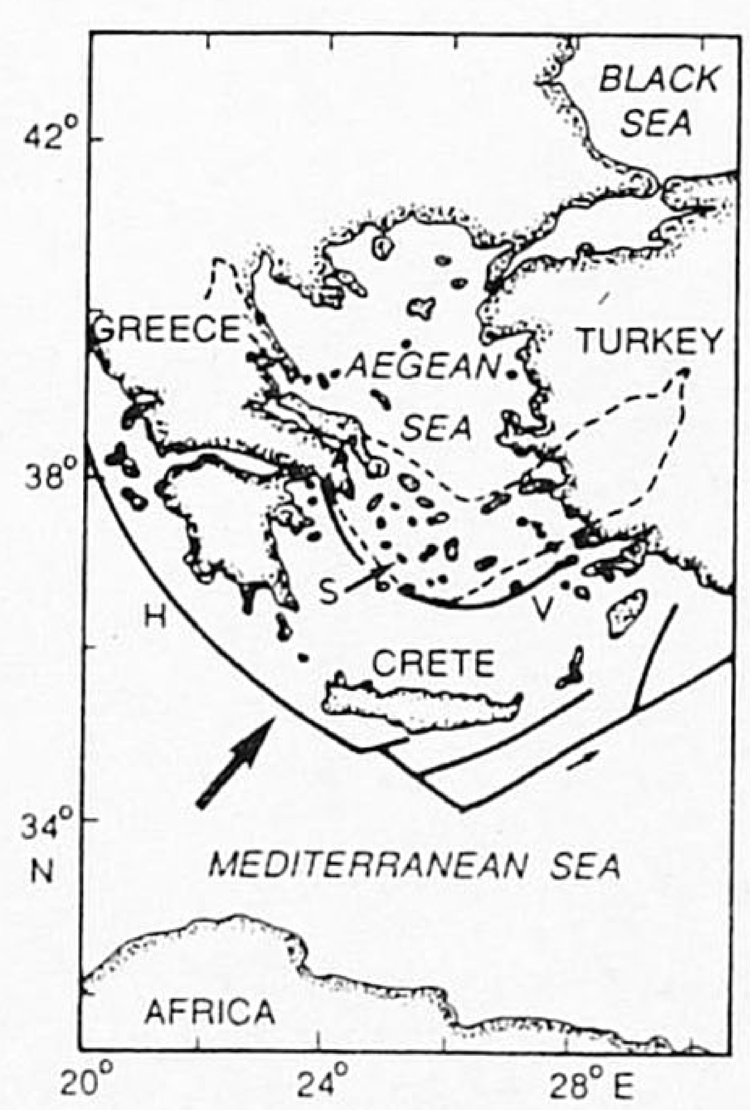

Figure 1. Tectonic sketch map of eastern Mediterranean Sea with location of Sifnos Island (S). General direction of subduction at Hellenic trench (H) is indicated (arrow). Volcanic arc (V) is subparallel to trench system. Dashed line encloses area with high-*P* metamorphism.

by parallel "fault" surfaces with constant vertical separation. Simple shear is simulated by imposing constant horizontal velocity within domains, following a kinematic technique described by van Wees et al. (1992). The method accommodates "bending" of domains at hinge lines by vertical shear within grid cells and also allows for readjusting the lithosphere-asthenosphere boundary according to the prescribed tectonic motions.

The temperature evolution at a subduction zone resulting from two-dimensional heat conduction is calculated taking into account advection terms arising from the imposed velocity field. Frictional heating is neglected, as the model invokes blocks of supracrustal affinity that have low internal strength at high temperatures (cf. strength envelope in Fig. 3). Vertically displaced material removed at the grid surface simulates erosion. The left (northern) grid margin is kept stationary; oceanic lithosphere is added at the other vertical boundary. Initial temperatures (Fig. 3, 65 Ma) in the oceanic lithosphere are assumed to be typical of a moderately mature basin. (Sensitivity analysis has shown that the initial oceanic geotherm is a relatively unimportant parameter in calculating the *P-T-t* evolution of subducted materials compared to rate of subduction.) Initial temperatures for continental lithosphere (Fig. 3, 65 Ma) are derived from a steady-state geotherm (Pollack and Chapman, 1977; Chapman, 1986) with a fixed surface heat flow of 80 mW/m^2 (Vitorello and Pollack, 1980). At the upper surface, $T = 0$ °C; heat flux perpendicular to the base of the lithosphere, defined as the $T = 1325$ °C isotherm, is assumed to be constant (Van den Beukel and Wortel, 1988). Heat production is assumed to be negligible in the mantle, 0.5 $mW \cdot m^{-3}$ in the lower crust, and variable in the upper crust such that it accounts for 40% of the surface heat flow. Conductivities for crustal and mantle rocks are 2.6 and 3.1 $W \cdot m^{-1} \cdot K^{-1}$ respectively.

Subduction of continental lithosphere occurs until buoyancy forces exceed its strength. Lithologic stratification implies the presence of rheologically weak zones in the lithosphere (e.g., Stephenson and Cloetingh, 1991), such as the boundary between upper and lower crust (see strength envelope in Fig. 3), favorable to the development of detachment faults. We envisage repeated delamination of lighter supracrustal layers along such weak zones. This process leads to the migration of detached slabs from the footwall of the subduction zone to its hanging wall (Davy and Gillet, 1986; Van den Beukel, 1990) and allows buoyancy-driven exhumation of a previously subducted domain. The denser microcontinental lower crust and mantle lithosphere meanwhile continue to subduct, eventually followed by resumed subduction of oceanic lithosphere.

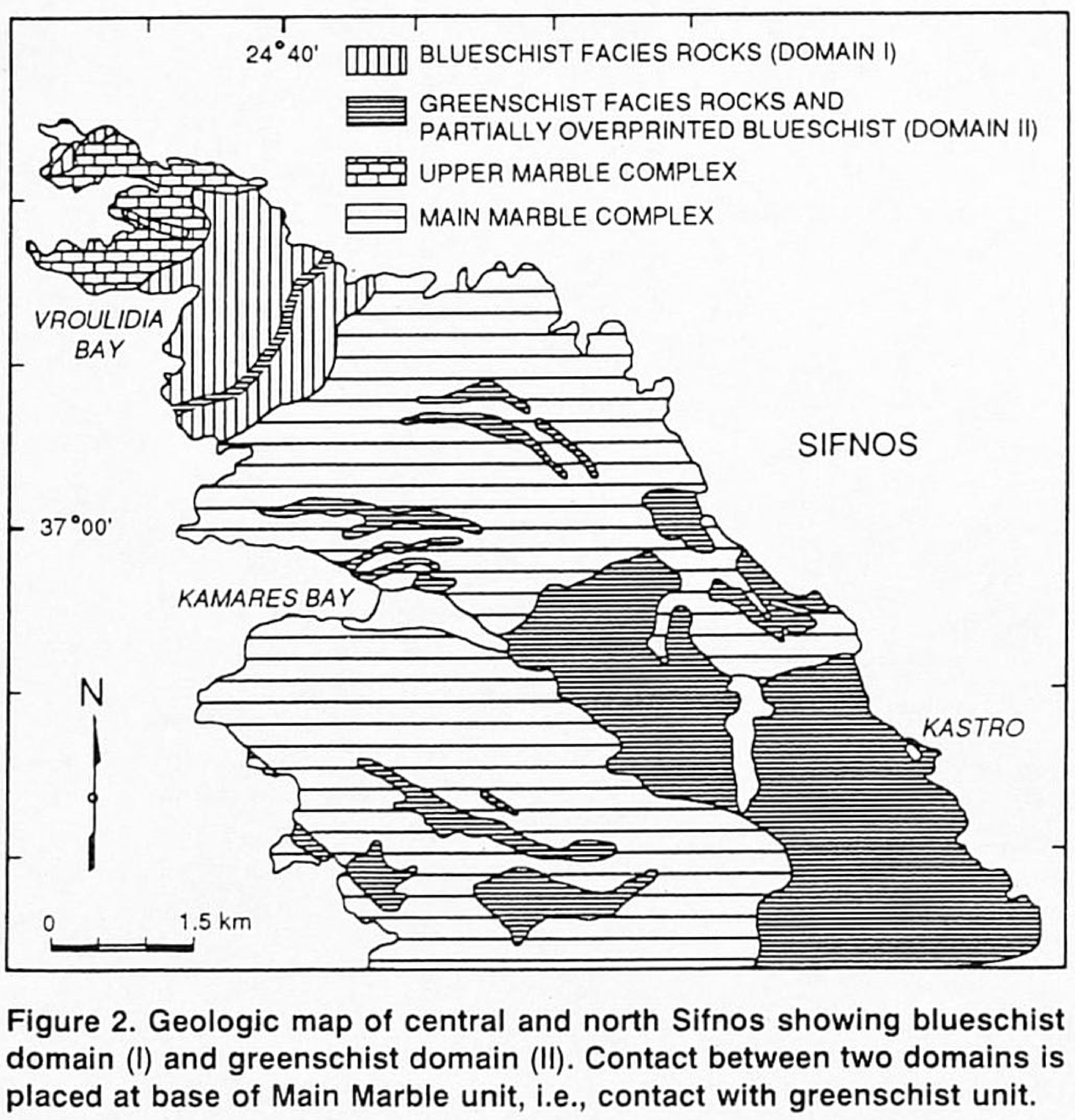

Figure 2. Geologic map of central and north Sifnos showing blueschist domain (I) and greenschist domain (II). Contact between two domains is placed at base of Main Marble unit, i.e., contact with greenschist unit.

The initial model geometry (Fig. 3, 65 Ma) represents the Late Cretaceous European margin and the colliding Cycladic microplate, separated by an incipient plane of subduction (dashed line), with the Mediterranean Sea lying to the south. The "incipient" subduction zone has been active previously, leading to oceanic closure at this time. The preexisting subduction history has little effect on the subsequent Tertiary *P-T-t* evolution. We adopt a subduction angle of 40° and a horizontal speed of 2.5 $cm \cdot yr^{-1}$, the latter corresponding to estimates for plate-convergence rates from Angelier et al. (1982). These parameters will cause subduction and detachment to occur in cycles of ~2.5 m.y. The initial locations of supracrustal protoliths for which we have calculated *P-T-t* pathways, corresponding to Sifnos domains I and II, are shown in Figure 3 as a triangle and a square, respectively.

The first two subduction-detachment cycles occurred in the 65–60 Ma interval, as illustrated in Figure 3. The second detached layer is juxtaposed beneath the first and remains closer to the locus of ongoing subduction (60 Ma). Its calculated *P-T* trajectory is one of monotonously increasing pressure and temperature. At 60 Ma, it is in turn detached and begins its buoyancy-driven ascent and decompression. The adopted exhumation rate is about 1.5 $mm \cdot yr^{-1}$, which is a high value for erosion-controlled uplift. Synuplift conductive and radiogenic heating leads to pressures and temperatures compatible with those inferred for peak metamorphic conditions of Sifnos rocks (box A) at about 57 Ma. The solid line in Figure 3 tracks the further *P-T* evolution of these strata during continued rapid exhumation, but in the absence of any other active cooling mechanism. This trajectory is clearly incompatible with the observations from Sifnos. In contrast, the other plotted *P-T* trajectories (i.e., short-dashed and long-dashed lines) are compatible with the Sifnos data. These cooling paths reflect the advective thermal effects of a nearby, colder subducted layer. They are strongly dependent on the rate of subduction and also on distance to the subducting layer.

A small amount of simple shear in the detached layer, internally consistent with the model kinematics and the shear couple between the adjacent rising and subducting domains, induces sufficiently different rates of uplift at each calculation point to account for a 10 m.y. difference in their cooling to 360 °C (phengite closure temperature). As a result (Fig. 3, 30 Ma), the position of Sifnos domain I (triangle) is 15 km shallower than domain II (square). After 30 Ma, we assume that active back-arc extension begins, leading to crustal thinning and back-arc sedimentary basin formation. In terms of model kinematics, the absolute motion of the hanging-wall layers at this time is one of uplift. However, relative displacements between layers change from reverse to normal as extension proceeds as the result of slab rollback (Angelier et al., 1982).

Continued relaxation of the still-depressed geotherm after 30 Ma is not sufficient, on the basis of our calculations, to cause greenschist overprinting in the time available (at or before 19 Ma). However, the influence of advective heating due to nonhomogeneous, back-arc, simple-shear extension (e.g., Wernicke, 1981) provides an alternative mechanism. Given a distance of 4 km (representing a mean value) below the extensional detachment, strata located in Sifnos domain II (square) are heated and reach greenschist conditions after a finite displacement of 50 km (Fig. 3, 22 Ma). The relation between overall amount of extension and degree of heating at any given point depends strongly on the distance from that point to the extensional detachment.

Subsequently, extension migrates to higher levels of the heated and weakened hanging wall, reactivating the old shear zones and leading to transposition of both calculation points to their present adjacent locations. During this time (22–18 Ma), model strata representing Sifnos domain I (triangle) are heated to 330 °C, and those representing Sifnos domain II (square) are cooled and pass through the 360 °C isotherm at 18 Ma (Fig. 3, 18 Ma), consistent with the observed isotope data.

IMPLICATIONS FOR HIGH-*P*–LOW-*T* TERRANES

Temperatures and pressures inferred from high-*P*–low-*T* metamorphism in Sifnos also represent the peak *P-T* conditions reached (Fig. 3, box A). The model results thus imply the action of a cooling mechanism that affected these rocks, in addition to exhumation. High exhumation rates alone would result in relatively low maximum temperatures, but, because of thermal inertia, these temperatures would be reached at lower pressures (solid line in the *P-T* diagram, Fig. 3). Uplift under the thermal influence of a nearby, colder, downgoing slab causes a temperature maximum to be reached early, at higher pressures. Similar processes of accretion, exhumation, and advective cooling for low-peak metamorphic temperatures have been proposed in the Alps (Rubie, 1984; Davy and Gillet, 1986). In the model here, the subducting layer that provides cooling is of oceanic affinity. Juxtaposition of additional continental domains would cause rocks at the calculation points to follow thermal paths (sub)parallel to the solid line, resulting in an inversion of the thermal structure with strata represented by the triangle being thermally overprinted instead of those represented by the square.

Both Sifnos structural domains display similar lithologies and evidence for similar high-*P* metamorphism. However, the isotopic data reveal substantially older cooling ages for domain I than for domain II. Assuming that the subduction and concomitant imposition of peak metamorphic conditions in both domains were roughly coincident in time, then the cooling ages mean that domain I was initially uplifted at a faster rate than domain II (Wijbrans et al., 1990). Incorporation of excess Ar or recrystallization effects cannot easily explain the tight clustering in isotopic ages in both domains. Exhumation of both domains in the vicinity of the actual subduction zone, which is required for the preservation of high-*P*–low-*T* metamorphic facies mineral assemblages, implies an exhumation process with the dominant translation occurring along shear zones parallel to the plane of continuing subduction. The calculations show that exhumation in the vicinity of the downgoing slab is critical for the preservation of blueschists.

Greenschist overprinting with *P-T* conditions corresponding to box B in Figure 3 is observed in the rocks of domain II but not in those of domain I (Schliestedt et al., 1987). This metamorphism occurred after the main cooling and exhumation stage, later than ~30 Ma as inferred from K-Ar closure ages of phengites, but prior to ~18 Ma as inferred from the K-Ar age of a greenschist mica. Assuming similar paleogeotherms for both domains during their *P-T-t* histories, these ages imply that not only was domain I uplifted at a faster rate than domain II, but that it had also reached a significantly higher crustal level during the 30–18 Ma interval. Differential movement between the two domains of Sifnos along the base of the Main Marble unit (Fig. 2) has been proposed by D. Avigad (1991, personal commun.). We infer that greenschist metamorphism is related to the widespread late Oligocene–early Miocene extensional phase in evidence throughout the Attic Cycladic blueschist belt (Spakman et al., 1988). It can be explained by the advection of hot, more deeply buried strata along north-dipping, simple-shear, extensional detachment zones (e.g., van Wees et al., 1992), as have been observed on the nearby island of Naxos (Buick, 1991). The surface underlying Sifnos domain II could be one such detachment, providing a means of juxtaposing underlying strata hot enough to cause the greenschist overprint during extension.

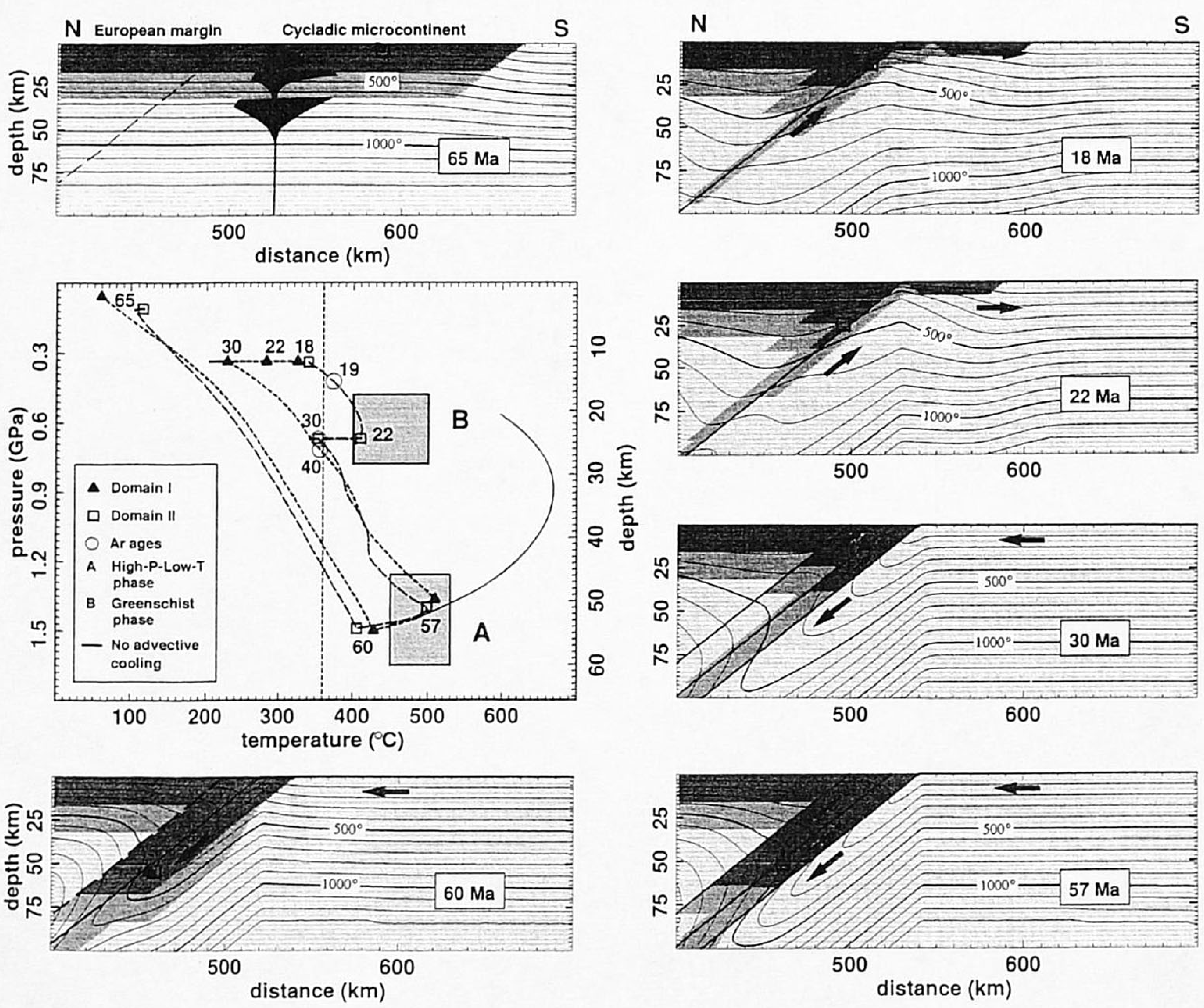

Figure 3. Model *P-T-t* evolution (dashed lines) for Sifnos domains I (solid triangles) and II (open squares) and corresponding thermal model geometries at selected times (for central part of finite-difference grid). Strength envelope calculated for extensional and contractional strength of a model lithosphere consisting of quartzite upper crust, diorite lower crust, and dunite mantle shows rheological stratification of crust (model and parameters after Stephenson and Cloetingh, 1991). Open circles indicate phengite K-Ar ages. Shaded boxes indicate peak metamorphic conditions: blueschist metamorphism (box A) and greenschist metamorphism (box B) (Schliestedt et al., 1987). Vertical dashed line in *P-T* diagram signifies 360 °C phengite closure temperature. In cross sections, upper and lower crust and mantle are shaded in dark, medium, and light gray, respectively. Solid arrows indicate block motions; heavy solid and dashed lines indicate faults related to microcontinent subduction. Light solid lines indicate calculated temperatures, contoured in 100 °C intervals.

CONCLUSIONS

Petrologic evidence for subduction to >50 km depth of blueschist- and eclogite-facies rocks of supracrustal affinity in the Attic Cycladic blueschist belt can be reconciled by a model in which advective cooling during exhumation prevented extensive static overprinting. Numerical results demonstrate that uplift alone is not a sufficiently effective cooling mechanism and that exhumation of such rocks must take place within 5–10 km of the active subduction zone to preserve blueschists and eclogites. Juxtaposition of additional supracrustal layers may have occurred in the central Cyclades. Repeated delamination of lighter supracrustal layers along a rheologically weak zone such as the boundary between the upper and lower crust leads to the migration of detached slabs from the footwall of the subduction zone to its hanging wall (Davy and Gillet, 1986; Van den Beukel, 1990) and allows buoyancy-driven exhumation of a previously subducted domain. Blueschist layers may have been repeatedly forced away from the subduction zone, resulting in local Miocene, upper amphibolite–facies metamorphism, accompanied by local crustal melting, heat advection by emplacement of I-type granitoids, and channeled fluid flow.

Neogene extension is a second key feature of the thermal evolution of the Attic Cycladic belt. The model demonstrates that rapid crustal extension along preexisting shear zones can account for local greenschist alteration, as is widely observed. By reactivating these shear zones, formed during the collisional stage, the model explicitly links early contractional tectonics with Neogene extension and back-arc basin development. It is in agreement with seismological and petrologic data that suggest continued convergence along the southern margin of the Eurasian plate since the late Mesozoic (Spakman et al., 1988; Wijbrans and McDougall, 1988; Schermer et al., 1990). Thus, the *P-T-t* evolution of Sifnos serves as an example for the tectonics of convergent margins in the eastern Mediterranean.

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