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Thrust sequences in the central part of the External Hellenides

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Abstract – The model of a foreland propagating sequence already presented for the External Hellenides is significantly modified in this paper. New data are used, including structural maps, cross-sections, stratigraphic determinations and seismic profiles. In general, thrusts formed a foreland propagating sequence but they acted simultaneously for a long period of time. Thus, during the Middle Eocene the Pindos thrust resulted in the formation of the Ionian–Gavrovo foreland and acted in tandem with the newly formed Gavrovo thrust within the basin until the Late Oligocene. The Gavrovo thrust consists of segments, showing that out-of-sequence thrusting was important. Thrust nucleation and propagation history is strongly influenced by normal faults formed in the forebulge region of the Ionian–Gavrovo foreland basin. Shortening rates within the Gavrovo–Ionian foreland are low, about 1 mm/year. Although thrust load played an important role in the formation of this basin, the additional load of 3500 m thick clastics in the basin enhanced subsidence and underthrusting.

Keywords: thrust faults, fold and thrust belts, propagation, Hellenides.

1. Introduction

The classic concept of polarity of mountain belts, which was originally established in the External Hellenides by Aubouin (1959, 1963), recognized the important role that the migration of the locus of thrusts played in the deformation history of external parts of an orogen. Mesozoic basins and ridges in the External Hellenides called 'isopic zones' were inverted during Tertiary times to form crustal-scale thrusts that carried rocks from east to west onto the Pre-Apulian autochthon (Fig. 1a). Aubouin's idea regarding orogenic polarity survived in the modern structural and stratigraphic studies of the External Hellenides, supported by the model of foreland propagating thrust faults (Brooks *et al*. 1988; Underhill, 1989; Clews, 1989; Jacobshagen, 1986). Although this model is accepted worldwide (Dahlstrom, 1970; Suppe, Chou & Hook, 1992), there are also orogenic belts that are characterized by out-of-sequence thrust movement (Eisenstadt & De Paor, 1987) or by synchronous thrust faults (Boyer, 1992; Price, 2001).

In this paper we present new stratigraphic data from syn-orogenic clastics to define the movement history of thrust faults in the area, and subsequently we try to integrate surface geological data derived from structural and sedimentological analysis with subsurface data from a seismic cross-section. We recognize a foreland propagating sequence of thrusting in which adjacent thrusts operated synchronously for a long time period.

2. Tectonic setting

The Hellenides are the result of the collision between the Apulian plate and the Eurasian continent (Dewey *et al*. 1973; Mountrakis, 1986; Doutsos *et al*. 1993). Collision began in Late Cretaceous–Early Eocene times with the obduction of the Pindos ophiolites onto the eastern Apulian margin (Fig. 1a) and continued throughout the Tertiary with westward (in present coordinates) prograding intracontinental deformation (Jones & Robertson, 1991). The end result is the so-called 'External Hellenides' fold-and-thrust belt consisting of a pre-orogenic and a syn-orogenic sedimentary sequence (Fig. 1b).

The pre-orogenic sequence is represented by Mesozoic carbonates. In Early Mesozoic times, shallow marine carbonates were deposited forming a platform. In Early–Middle Jurassic times, rifting subdivided this platform into ridges, the Gavrovo and Pre-Apulian zones, and basins, the Ionian and Pindos zones, filled with shallow marine carbonates and with pelagic carbonates and radiolarites, respectively (BP, 1971; Karakitsios, 1995). Post-rift sedimentation persisted throughout the Cretaceous period, leading to the deposition of shallow water carbonates on the ridges and deep-water carbonates in the basins.

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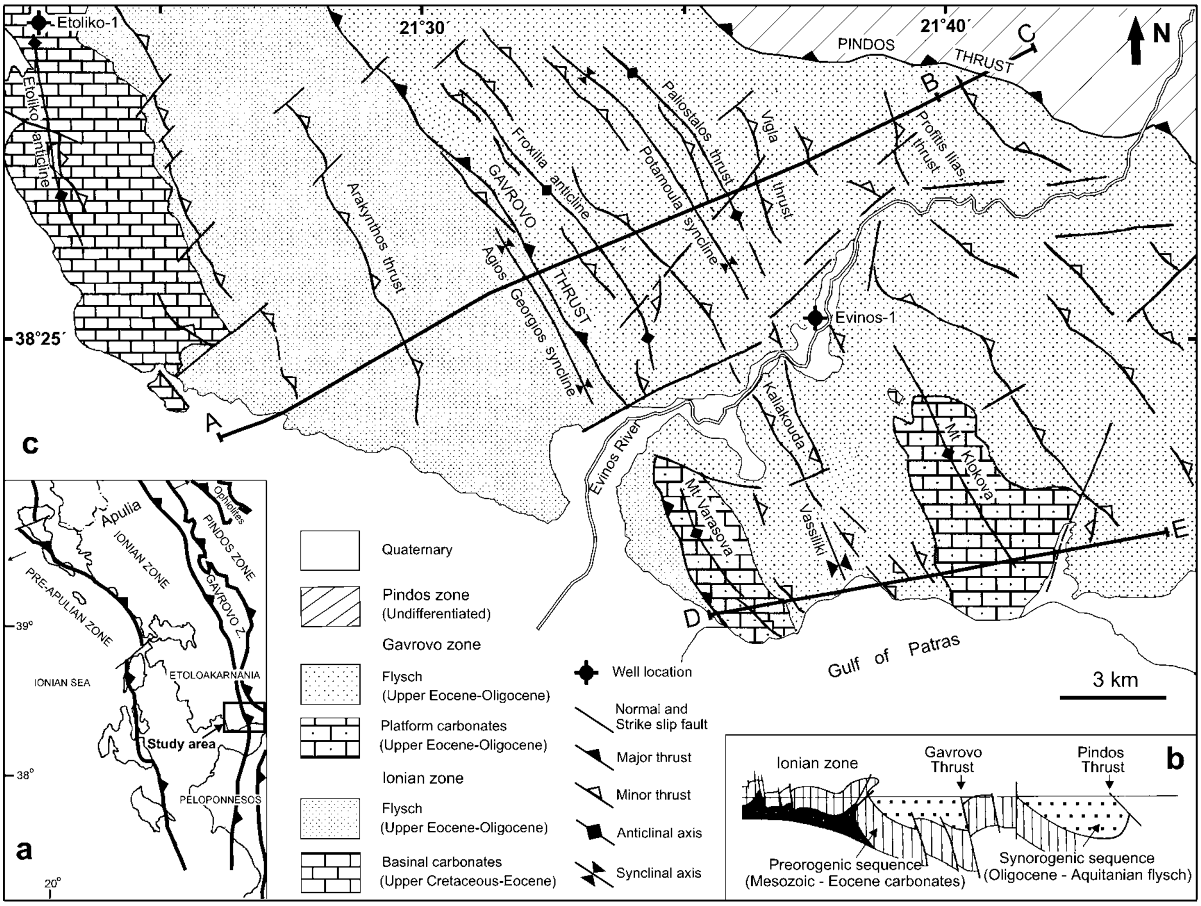

Figure 1. (a) Simplified map showing the main structural features of the External Hellenides. (b) Schematic tectonic cross-section of the External Hellenides (IFP, 1966). (c) Tectonic map of the study area showing structural elements. Cross-sections along lines A–C, D–E are shown in Figure 2. Seismic line along line A–B is shown in Figure 3.

The syn-orogenic sequence consists of flysch deposits accumulated diachronously across the Apulian platform (IFP, 1966). To the east a wedge-shaped prism with a maximum thickness of 2500 m was formed as ophiolites were obducted onto the eastern margin of the Apulian plate in Early Eocene times (Koch & Nicolaus, 1969; Skourlis & Doutsos, 2003). During Late Eocene times, this prism and the underlying pre-orogenic rocks of the Pindos zone detached from their ophiolitic basement and were emplaced along the Pindos thrust onto the Gavrovo zone (Fleury, 1980; Degnan & Robertson, 1998; Xypolias & Doutsos, 2000). The Pindos thrust remained active until Late Oligocene times (Fleury, 1980) and led to the formation of the Pindos cordillera which supplied erosional detritus westward to the Ionian–Gavrovo foreland (Aubouin, 1963; Richter, 1976; Faupl, Pavlopoulos & Migiros, 1998). The thrust load exerted by the Pindos cordillera has been considered to be responsible for the formation of this foreland basin (Clews, 1989). A turbidite sequence up to 5 km thick was deposited in the basin over the interval between the Priabonian and Middle Miocene (Jenkins, 1972; Avramidis *et al*. 2002). These turbidites are considered to be submarine fan deltas showing transitions from coarse to fine lithofacies with increasing distance from the front of the advancing Pindos thrust (Piper, Panagos & Pe, 1978).

The study area occupies the eastern part of the Ionian–Gavrovo foreland. As presented in Figure 1b, previous structural interpretations in this area considered the boundary between Ionian and Gavrovo zones to have been mildy inverted during the Late Oligocene to form the Gavrovo thrust (Aubouin, 1959; Brooks *et al*. 1988; Dercourt & Thiebault, 1979; IFP, 1966).

In this paper we provide evidence that the Gavrovo thrust is a crustal-scale thrust with displacement of more than 10 km and that it was active throughout the Oligocene simultaneously with the Pindos thrust.

3. Tectonostratigraphy

Previous and new stratigraphic data are used to describe vertical movements in the Ionian–Gavrovo foreland. These include two successive stages: an Early Tertiary flexural uplift followed by Oligocene flexural subsidence.

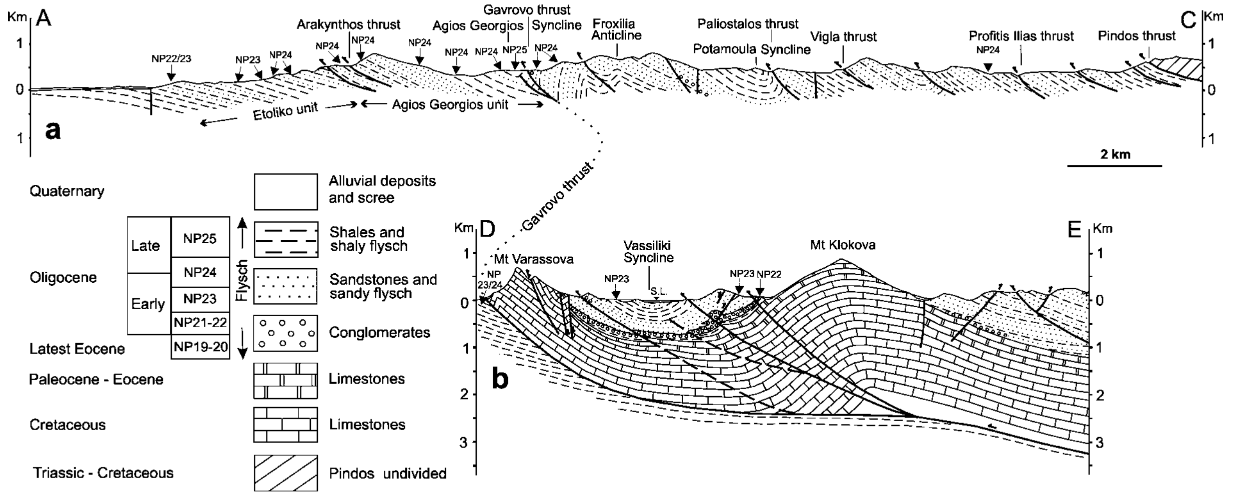

Figure 2. Tectonic cross-sections through the Ionian and Gavrovo zones in the Etoloakarnania area. Location of sections shown in Figure 1.

3.a. Flexural uplift

In Late Mesozoic times, the Apulian platform in Etoloakarnania (Fig. 1a) was composed of shallow water carbonates in the Gavrovo zone that passed gradually to the west into post–rift pelagic carbonates of the Ionian zone (Karakitsios, 1995). Later, within the Paleocene–Eocene time interval, deep-sea sedimentation persisted within the Ionian zone, whereas in the Gavrovo zone, several interruptions of the sedimentation have been reported (Jenkins, 1972). This time interval is represented in the western part of the Gavrovo zone on Mt Varassova by a few metres thickness of condensed and brecciated carbonates as well as by 350 m thick shallow marine carbonates at the central part of this zone on Mt Klokova (Figs 1c, 2b, cross-section D–E). Upper limestone members on Mt Klokova are intercalated with thin bauxite horizons indicating a sub-aerial exposure of the area during Middle Eocene times (BP, 1971; Fleury, 1980).

Since this Early Tertiary uplift and erosion in the Gavrovo zone took place simultaneously with the obduction of the ophiolites further east at the eastern margin of the Apulian plate (Doutsos *et al.* 1993), the Gavrovo zone may represent a flexural bulge induced by bending at the front of the obduction. Several syn-sedimentary faults identified within the Eocene limestones of the Ionian zone can be attributed to the same mechanism (see also Fig. 3).

3.b. Flexural subsidence

In the Late Eocene, the Pindos thrust load resulted in a flexural subsidence in the Ionian–Gavrovo foreland, in which thick syn-orogenic clastics were accommodated (Clews, 1989). About 60 samples collected from the foreland basin clastic sequences along the cross-sections of Figure 2 have been dated by nannofossil biostratigraphy based on the biozonal scheme of Martini (1971) and incorporated into the Integrated Magnetobiostratigraphic Scale (IMBS) of Berggren *et al.* (1995). For nannofossil analysis, smear slides have been prepared with standard techniques and studied under light microscopes. In order to search thoroughly for the marker species in flysch deposits, around 1650 fields of view have been investigated per slide, under 1250×.

According to the tectonostratigraphic evolution in the foreland, two basins have been identified, the Ionian and the Gavrovo basin.

3.b.1. The Ionian basin

The transition from carbonatic to clastic sedimentation took place in the Priabonian (Late Eocene) times (NP19–NP20) (*sensu* Martini, 1971; Berggren *et al.* 1995) as indicated by nannofossil markers (*Discoaster barbadiensis*). The overlying clastics can be subdivided into the Etoliko and the Agios Georgios units deposited at the front of the Arakynthos and Gavrovo thrust, respectively (Fig. 2, cross-section A–C). The Etoliko unit comprises thin-bedded turbidites up to 1700 m thick which consist of shaly flysch at the bottom and a more sandy flysch at the top. Sedimentation began during Early Oligocene times (NP23), as indicated by the presence of *Sphenolithus predistentus* and the absence of *Reticulofenestra umbilicus*, and lasted until Middle Oligocene times (NP24), as shown by the presence of *Sphenolithus distentus* and *Sphenolithus ciperoensis*. The Agios Georgios unit is up to 1500 m thick and consists of massive sandstones and conglomerates at the bottom and marly siltstones with intercalations of thin-grained sandstones and slumps at the top. Sedimentation began in Middle Oligocene times (NP24) and lasted until Late Oligocene times (NP25),

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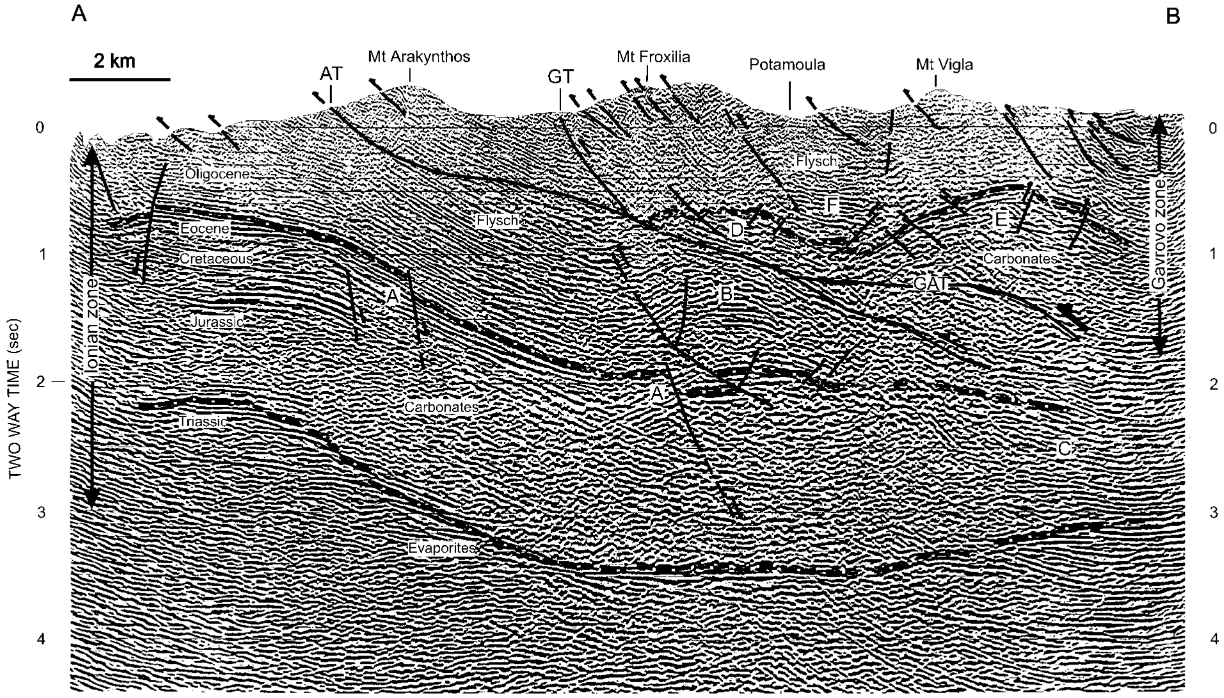

Figure 3. Seismic line (A–B) through the Ionian and Gavrovo zones in the Etoloakarnania showing the Gavrovo–Arakynthos thrust trajectory. A: normal faults; B: structural culmination; C: thinning of the Ionian carbonate sequence; D: Varassova anticline; E: Klokova anticline; F: thinning of syn-orogenic clastics toward the limbs of the Potamoula syncline; AT: Arakynthos thrust; GT: Gavrovo thrust; GAT: Gavrovo–Arakynthos thrust. The location of the seismic line is given in Figure 1.

the latter defined by the contemporaneous presence of *Sphenolithus ciperoensis* and *Reticulofenestra bisecta*.

Based on bed thickness, sandstone/mudstone ratio and the presence of internal structures, the turbidites of both units are described in terms of inner fan, middle fan and outer fan (*sensu* Ricci Lucchi, 1975). The Etoliko unit represents a retrogradational sequence (*sensu* Howell & Normark, 1982) while the Agios Georgios unit corresponds to a progradational sequence of middle fan associations in the bottom, graded upwards to inner fan associations.

3.b.2. The Gavrovo basin

The clastic sedimentation within the Gavrovo started at the Eocene–Oligocene boundary with the deposition of marls, thin-bedded sandstones and conglomerates. On Mt Varassova these conglomerates contain pebbles of the underlying basement (Fleury, 1980), suggesting that at least some parts of the Gavrovo zone were exposed at the Earth's surface during the deposition of the conglomerates. Furthermore, there are several places on Mt Klokova where clastic deposits unconformably overlie the pre-orogenic basement (Richter & Mariolakos, 1975). Up to 1800 m of syn-orogenic clastics were accumulated within the Gavrovo basin during Oligocene times (Fig. 2, cross-sections A–C and D–E). They are composed of inner fan to slope deposits at the basal part of the basin, graded upwards to middle to outer fan deposits.

4. Subsurface data

For oil exploration purposes, several seismic reflection profiles have been acquired in the study area. The profile shown in Figure 3 crosses the Ionian–Gavrovo foreland basin in an ENE–WSW direction, perpendicular to the structural grain of the External Hellenides.

4.a. Geometry

A crustal-scale thrust trajectory, 'the Gavrovo–Arakynthos Thrust system', separates the seismic profile in Figure 3 into two structural domains, one in the footwall and one in the hanging wall of the thrust.

The structural domain in the footwall is characterized by an E-dipping monocline made of about 4 km thick Mesozoic carbonates of the Ionian zone buried under 3.5 km thick syn-orogenic clastic sediments (Fig. 3). The subsurface stratigraphy of the Ionian and Gavrovo zones along the seismic profile has been correlated with data from two wells, Etoliko-1 and Evinos-1, respectively (Fig. 1c). The top of the carbonatic sediments is strongly affected by steeply dipping normal faults having displacements up to 500 m (Fig. 3, points A). These extensional structures often occur at the deeper parts of flexural basins as in the case of the Alpine molasse basins (Coward, 1994), as well as in the adjacent foreland basin of the northwest Peloponnesos (Kamberis *et al*. 2000). These faults are probably related to bending deformation within the forebulge

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area. Similar processes are identified as driving the initiation of normal faults in the Pre-Apulian Zone (Underhill, 1989). Although a reactivation of these faults during the subsequent syn-orogenic contraction cannot be identified in the seismic profile, they often acted as a stress riser leading to the formation of a structural culmination (Fig. 3, point B). This type of culmination is composed of forward and backward listric thrust faults.

Pre-orogenic sediments preserved a nearly constant thickness in the west and central parts of the monocline but became thinner eastwards towards the Gavrovo carbonate platform (Fig. 3, point C). This thinning leads us to assume that the carbonates at the easternmost end of the Gavrovo–Arakynthos thrust are rooted in an area which was previously affected by extensive rifting, very probably in the Middle Jurassic period.

The lower part of the syn-orogenic prism in the footwall of the Gavrovo–Arakynthos thrust is mainly of Early Oligocene age (NP23), is about 1700 m thick and shows moderate deformation by fault bend folds.

The structural domain in the hanging wall of the Gavrovo–Arakynthos thrust is strongly affected by a ramp-flat geometry. Two low-amplitude structural culminations were formed above the ramps in the eastern part of this thrust (Fig. 3, points D and E). These two ramp anticlines, corresponding to the Varassova and Klokova anticlines mapped further south (Fig. 1c), are separated by a ramp syncline, the Potamoula syncline (Fig. 3, point F). At the frontal area of the Varassova anticline, the Gavrovo–Arakynthos thrust splays into two segments. The easternmost segment, the Gavrovo thrust, is short and strongly curved, breaks across the frontal limb of the anticline, and is accompanied by a dense pattern of imbricate stacks. The westernmost segment, the Arakynthos thrust, is long and moderately curved and is associated with low amplitude fault-bend folds and an imbricate stack consisting of closely spaced thrusts. The thickness of the syn-orogenic clastics in the hanging wall of the Gavrovo–Arakynthos thrust is up to 1800 m (Agios Georgios unit).

4.b. Kinematics

Seismic and surface data has been used to reconstruct the kinematic history of the Gavrovo–Arakynthos thrust and the Pindos thrust.

The basal part of the flysch sediments within the Vassiliki syncline (between Mt Klokova and Mt Varassova) was deposited during Early Oligocene times (Fig. 2, cross-section D–E) and shows a pronounced thinning toward the limbs of the adjacent anticlines, indicating that these structures are growth folds (Fig. 3, point F). Hence, double ramp thrusting below these anticlines must have been initiated during Early Oligocene times.

Later, this thrust trajectory must have propagated westward to form the Arakynthos thrust repeating the Middle Oligocene (NP24) sequence, as shown in Figure 2, cross-section A–C. After this event, during Late Oligocene times (NP25), the Arakynthos thrust lay dormant and clastic deposition migrated eastwards into the Agios Georgios syncline (Fig. 2, cross-section A–C). This syncline developed on the footwall of the newly formed Gavrovo thrust segment.

The above described out-of-sequence thrusting may have been generated by gradual amplification and uplift of the underlying structural culmination at the eastern part of the monocline (Fig. 3, point B). It may have caused the deactivation of the Arakynthos thrust and produced a stress concentration (*sensu* Wiltschko & Eastman, 1983) which favoured the formation of the Gavrovo thrust ramp.

Although the seismic profile in Figure 3 does not reach the Pindos thrust, some characteristics of the final kinematic stages of this thrust can be inferred based on the correlations of surface and subsurface data. The Pindos thrust was active until Late Oligocene times, as evidenced by the occurrence of nannofossil markers of biozone NP24 within the Gavrovo basin, even in front of the Pindos thrust. These stratigraphic determinations confirm previous work done in this area (Fleury, 1980). From the proximal (east) to the distal (west) part of the basin, the thickness of the sediments decreased from 2200 m to 1600 m. As can be seen in Figure 3, this stratal thinning took place abruptly at the backlimb of the Klokova anticline, which uplifted at the hanging wall of the Gavrovo–Arakynthos thrust. Hence the Gavrovo basin was formed as a thrust-sheet top basin (or piggy-back basin) in relation to the newly formed Gavrovo–Arakynthos thrust, while at the same time the older Pindos thrust remained active in its proximal part.

5. Synthesis

Emplacement of the Pindos nappe onto the eastern parts of the Apulian plate took place in the time interval between Late Eocene and Late Oligocene times (Fig. 4) and therefore Oligocene clastic sediments in the Ionian–Gavrovo foreland record only the late stages of Pindos thrust activity. Synchronous with this activity, a newly forming thrust, the Gavrovo–Arakynthos thrust, subdivided the Ionian–Gavrovo foreland basin into two sub-basins (Fig. 4b). The nucleation of this thrust took place in the Ionian–Gavrovo transition, which consisted of carbonates thinned and rifted during Middle Jurassic times (Fig. 4a).

The Gavrovo–Arakynthos thrust represents an out-of-sequence thrust system whose western part splayed into two curvilinear segments: the older Arakynthos segment in the front (Fig. 4c) and the younger Gavrovo segment in the rear (Fig. 4d). The eastern part of the Gavrovo–Arakynthos thrust is located above

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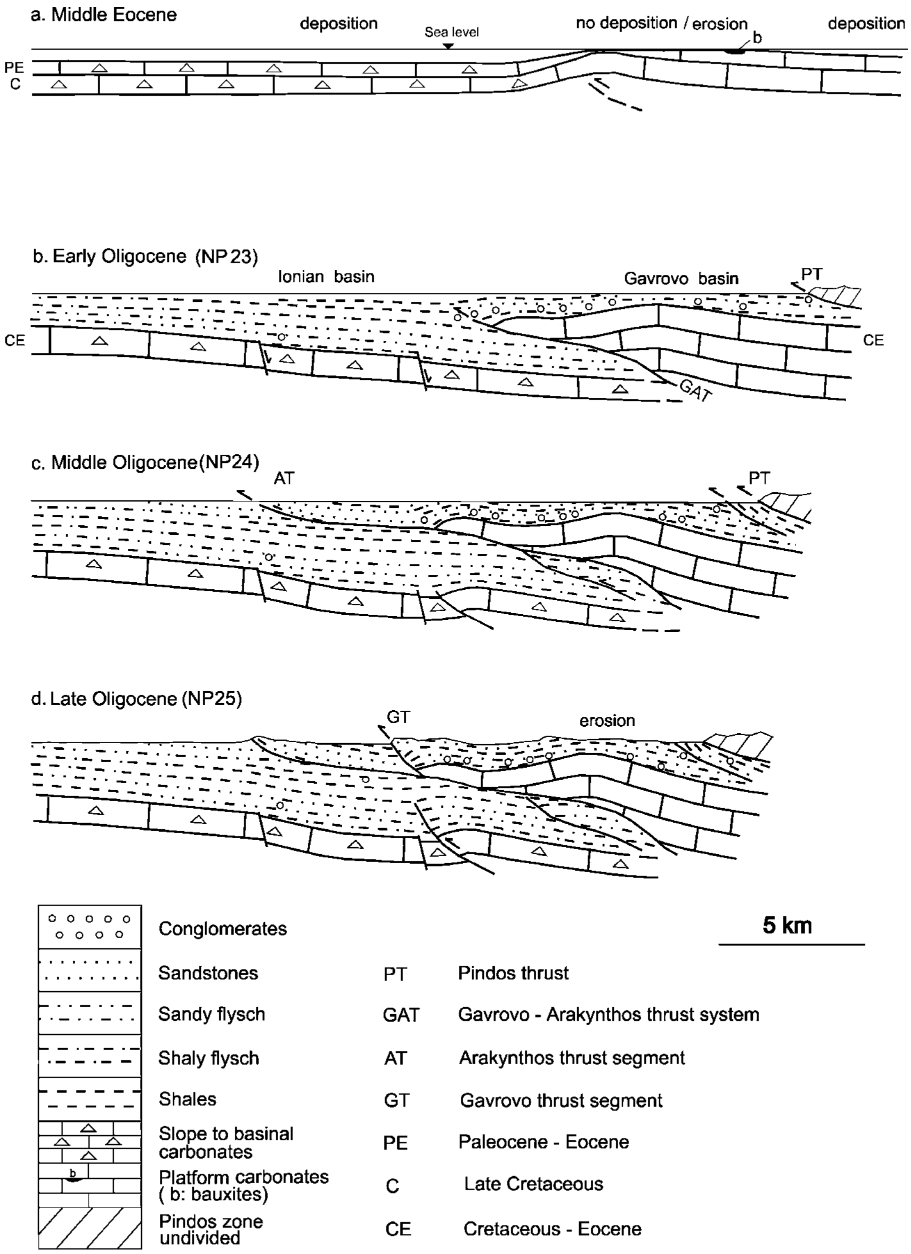

Figure 4. Schematic cross-sections are showing foreland evolution of the External Hellenides in the Etoloakarnania.

an inverted normal fault of the Ionian monocline and comprises two steps forming the Klokova and Varassova ramp anticlines. These are growth folds that remained below sea-level throughout Oligocene times, as uplift rates in this area are less than sedimentation rates.

As movements along the Gavrovo–Arakynthos thrust are on the order of at least 10 km and the internal deformation of the rocks was estimated to be 23 % (Robertson Research International, unpub. report, 1996) the resulting shortening rate during Oligocene times along the seismic profile illustrated in Figure 3 is about 1 mm/year. Much of this shortening must have taken place during Middle Oligocene times as the Arakynthos segment reached the Earth's surface and became inactive. However, it is noteworthy that the mean shortening rate of 1 mm/year is small compared to that of 5 mm/year reported from the Pindos zone (Skourlis & Doutsos, 2003). This decrease of shortening rates westward towards the Pre-Apulian autochthon is a common phenomenon taking place at the external parts of many orogens (see Apennines: Zoetemeijer *et al*. 1992).

Furthermore, some key points may be referred to in order to evaluate factors controlling foreland basin formation in the Etoloakarnania region. Clearly the thrust load exerted by the Pindos cordillera plays an important role in the evolution of the Gavrovo basin filled with up to 1800 m thick syn-orogenic clastics. However, this thrust load is insufficient to explain the

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accumulation of 3500 m thick syn-orogenic clastics in the Ionian basin, inasmuch as the hanging wall of the Gavrovo–Arakynthos thrust remained below sea-level throughout Oligocene times. Weakening of the crust during the Middle Jurassic rifting, affecting the Ionian zone, did not cause enhanced subsidence due to cooling of the crust (see also Thompson *et al*. 2001) in the long time interval between the Middle Jurassic rifting and the Late Eocene foreland evolution. The deep pelagic bathymetry that existed in the Ionian subbasin at the onset of clastic sedimentation (Fig. 4a; see also Gonzalez & Bonorino, 1996) may be responsible for the accumulation of these thick clastics that caused additional loading and foreland subsidence (see also Stockmal, Beaumont & Boutilier, 1986; Roure *et al*. 1994; Sandiford, 1999). Underthrusting commenced at the Gavrovo/Ionian boundary which very probably represented a Middle Jurassic normal fault within the Apulian margin.

6. Conclusions

New results from subsurface and field data demonstrate that the classic model of a foreland propagating sequence of thrusting for the External Hellenides needs significant modification. It can now be shown that two crustal-scale faults, the Pindos thrust and the Gavrovo–Arakynthos thrust, acted simultaneously until Late Oligocene times; the Gavrovo–Arakynthos thrust, with a complex propagation history involving out-of-sequence thrusting and flexural bending during the first stages of the syn-orogenic basin, induced local uplift and the formation of normal faults, which later, during orogenic shortening predetermined the site of thrust ramps and anticlines. Furthermore, the correlation of the stratigraphic, structural and sedimentological data show that shortening rates within the Gavrovo–Ionian foreland were low, about 1 mm/year, during the contractional episode.

Although the thrust load of the Pindos cordillera played an important role in formation of the Ionian–Gavrovo foreland basin, additional load of the accumulated clastics and superimposition on an inherited passive continental margin structure led to enhanced subsidence and underthrusting.

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