The Influence of Preexisting Structure and Halokinesis on Organic Matter Preservation and Thrust System Evolution in the Ionian Basin, Northwest Greece[1]

Vassilis Karakitsios[2]

ABSTRACT

The opening of the Ionian basin and its internal differentiation is attested to by lateral facies and thickness variations of the formations deposited during the Pliensbachian and Tithonian (synrift formations). The beginning of the synrift sequence is represented by the Siniais Limestones (Pliensbachian) and their lateral equivalent, the Louros Limestones. The geometric characteristics of the extensional basin depend on both extension related to the latest opening of the Tethys ocean and halokinesis of the Ionian zone evaporitic substratum. The orientation of the extensional regime is deduced from the direction of stratigraphic pinchouts of the synrift formations, directions of transport of slump features, and direction of movement on synsedimentary faults observed in the base of these formations in the half grabens. The postrift period is marked by an unconformity at the base of the Vigla Limestones, which represents the early Berriasian breakup during which sedimentation was synchronous in the entire Ionian basin.

The accumulation of organic matter in the "Lower and Upper Posidonia beds" of the Ionian zone during the Toarcian and Tithonian is directly related to the geometry of the synrift period of the Ionian basin. Restricted subbasins were formed where the geometry of the basin favored stagnation and consequently the locally euxinic conditions of bottom waters. Anoxic conditions persisted locally to the postrift period in areas where the "Upper Siliceous Zone" (Albian–Cenomanian) of the Vigla Limestones is well developed; these areas probably represent subbasins that were preserved by the continuation of halokinetic movements during the postrift period.

During the early Miocene Alpine orogeny, collision-related compressive stresses on the margin induced the reactivation of preexisting fractures, which were responsible for the inversion tectonics that affected the Mesozoic basin. The geometric characteristics of the inverted basin were dependent on lithology (evaporites), geometry of the extensional structures, and direction of the compressional phase. The observed geometries do not always correspond to the classical scheme of inversion tectonics. The geological evolution in the Ionian basin is an example of inversion tectonics of a basin with an evaporitic substratum.

The opening of the Ionian basin and the inversion tectonics influence both the source rocks and the probable hydrocarbon traps of the Ionian zone.

INTRODUCTION

It has long been recognized that the stratigraphic distinctiveness of individual thrust sheets in the Alpine system (the tectonostratigraphic units of classic Alpine stratigraphy) is related to the pretectonic paleogeography. This paleogeography is a reflection of Mesozoic extensional geometry associated with the extensional events that preceded the development of the Tethys ocean (Trümpy, 1980; Lemoine et al., 1986; de Graciansky et al., 1989). Inversion tectonics refers to the control exerted by early extensional faults on the compressional geometry developed during later orogenic deformation.

At the scale of hundreds of kilometers, the whole Alpine belt can be considered as the inverted margin of the Tethys ocean in response to the collision of Apulia against Europe (de Graciansky et al., 1989). On a smaller scale of a few tens of kilometers, the various subbasins of the Hellenic Tethyan margin have been inverted to produce the main Hellenic thrust sheets or folded zones. This occurred successively from inner (eastern)

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[2]Department of Geology, University of Athens, Panepistimioupoli, 15784 Athens, Greece.

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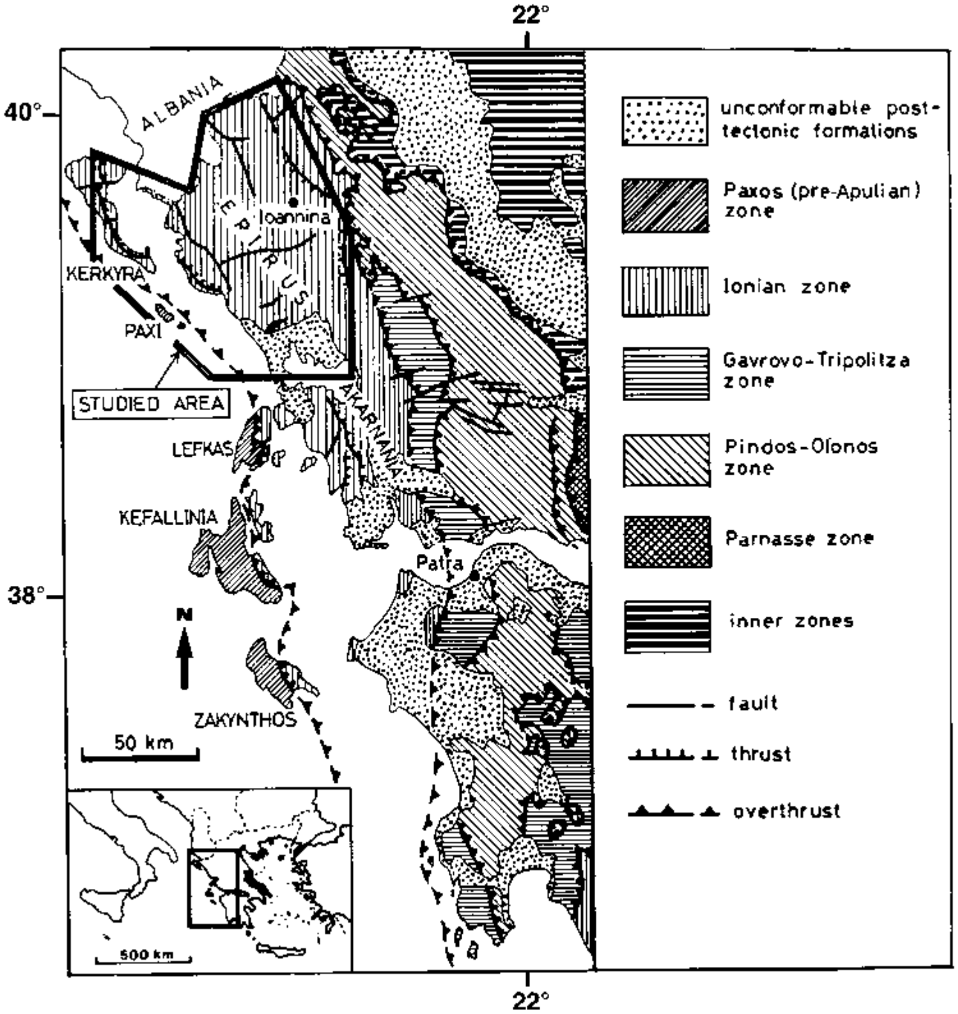


Figure 1—Structural map of western continental Greece.

zones to external (western) zones. One of the most representative examples of this inversion is the Ionian zone of the external Hellenides (Karakitsios, 1990, 1992).

The Ionian zone of northwest Greece (Epirus region) constitutes part of the most external zones of the Hellenides (Paxos zone, Ionian zone, Gavrovo-Tripolitza zone; Figure 1). The rocks of the Ionian zone range from evaporites (Triassic), through a varied series of carbonates and lesser cherts and shales (Jurassic through upper Eocene), followed by flysch (Oligocene) (Figure 2).

In the early Liassic, northwest Greece was covered by a huge carbonate platform (Bernoulli and Renz, 1970; Karakitsios, 1992). In the Pliensbachian, extensional stresses associated with the opening of the Tethys ocean caused the opening of the Ionian basin (Bernoulli and Renz, 1970; Karakitsios, 1990, 1992). Even though the production of platform carbonates persisted through the entire Jurassic in the adjacent Paxos (pre-Apulian) and Gavrovo-Tripolitza zones, the Ionian basin became an area of stronger subsidence and faulting. This paleogeographic configuration continued with minor off- and onlap movements along the basin margins until the late Eocene, when orogenic movements and flysch sedimentation began. In the Gavrovo-Tripolitza and Ionian zones the main orogenic movements took place at the end of the Burdigalian, whereas in the Paxos and Apulian zones they occurred during the Pliocene–Pleistocene (IGRS-IFP, 1966; Bizon, 1967).

This study will show

• that the evaporitic Ionian substratum "halokinesis" (Trusheim, 1957, 1960) influenced the intense listric block-faulting. The latter affected the early Liassic shallow platform (Pantokrator Limestones) and resulted in the formation of several small, structurally controlled basins

• that the geometric characteristics of these synrift sequences are evidenced by facies and thickness variations in the Pliensbachian to Tithonian

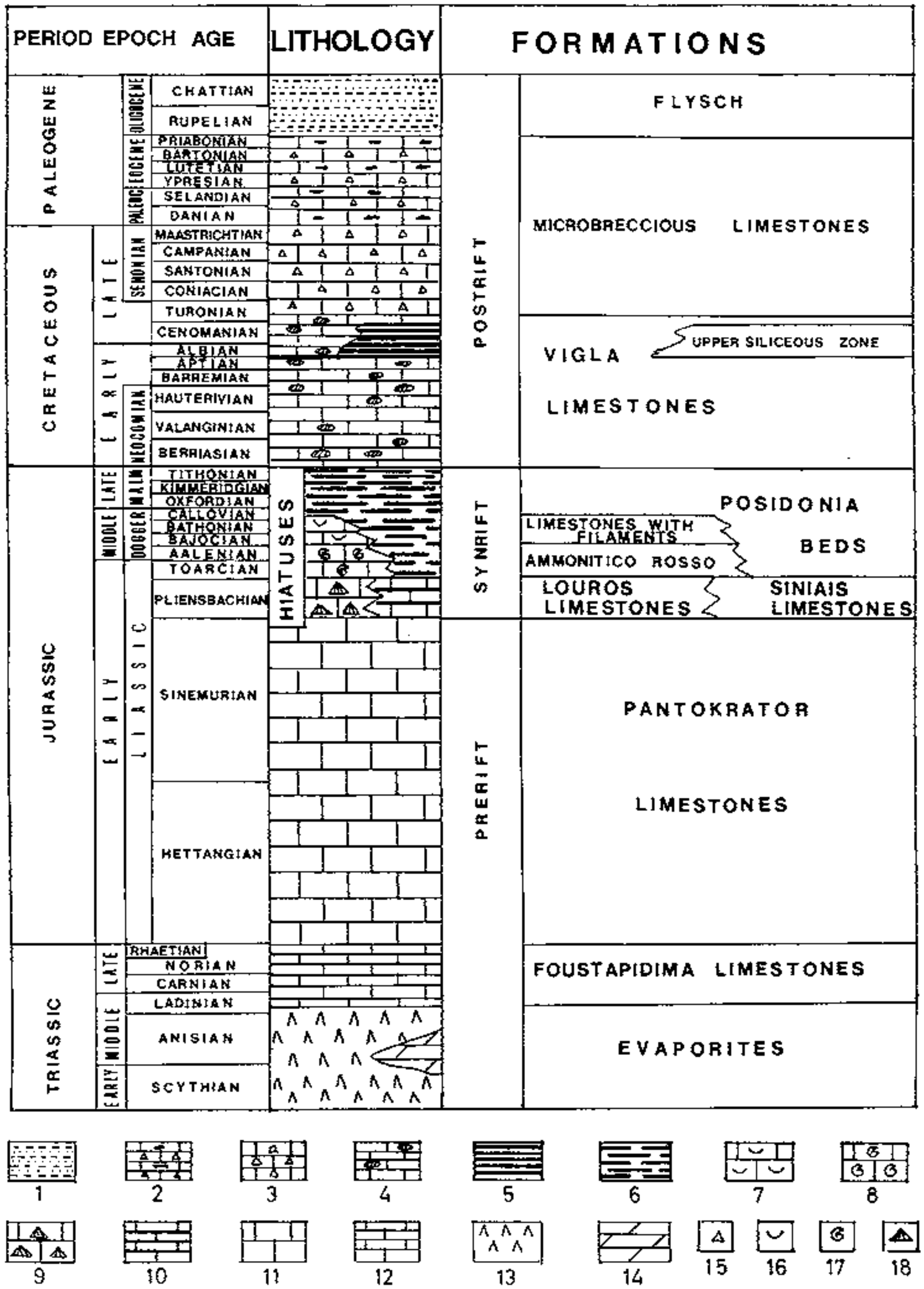


Figure 2—Representative stratigraphic column of the Ionian zone. 1 = pelites and sandstones; 2 = cherty limestone with clastic material; 3 = pelagic limestones with clastic material; 4 = pelagic cherty limestones; 5 = cherty beds with green and red clay, sometimes shaly; 6 = pelagic limestones, marls, and siliceous argillites; 7 = pelagic limestones with pelagic lamellibranches; 8 = pelagic, red, nodular limestones with ammonites; 9 = micritic limestones with small ammonites and brachiopods; 10 = pelagic limestones; 11 = platform limestones; 12 = platy black limestones; 13 = gypsum and salt; 14 = dolomites; 15 = breccia; 16 = section of pelagic lamellibranch (filament); 17 = ammonite; 18 = brachiopod.

deposits, and by the direction of synsedimentary features

• the control of basin geometry in organic matter preservation

• the reactivation of preexisting extensional fractures from Pliensbachian through Tithonian during compressional tectonics of Alpine orogeny (inversion tectonics of the basin)

• that the geometric characteristics of the inverted basin that depend on the lithology (evaporites) and the geometry of the extensional structures do not always correspond to the classical scheme of inversion tectonics.

GEOMETRY OF SYNRIFT SEQUENCES DEPOSITED DURING EXTENSIONAL FAULTING FROM PLIENSBACHIAN THROUGH TITHONIAN

Stratigraphy of extensional basins, in its simplest form, may exhibit three distinct sequences

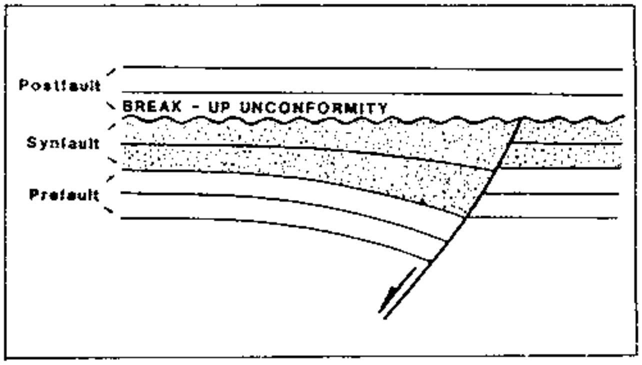


Figure 3—Schematic diagram of sediments accumulated before, during, and after extensional fault movement. The breakup unconformity is shown as a wavy line at the top of the synrift sequence (after Williams et al., 1989).

(Williams et al., 1989) (Figure 3): (1) a prerift sequence that was deposited prior to any extensional fault movement, (2) a synrift sequence deposited during extensional faulting, evidenced by marked stratigraphic thickness changes from footwall to hanging wall indicative of growth faulting, and (3) a postrift sequence deposited after the cessation of extensional faulting. The postrift sequence may be deposited after a period of nondeposition and/or erosion marked by a breakup unconformity where part of the synrift sequence has been removed.

In the Ionian zone, the prerift sequence is represented by the early Liassic Pantokrator Limestones (Aubouin, 1959; IGRS-IFP, 1966; Karakitsios, 1990, 1992). These shallow-water limestones overlie Early to Middle Triassic evaporites through Foustapidima Limestones of Ladinian–Rhetian age (Renz, 1955; Pomoni-Papaioannou and Tsaila-Monopolis, 1983; Dragastan et al., 1985; Karakitsios and Tsaila-Monopolis, 1990). The "subevaporite beds" of the Ionian zone in western Greece do not crop out, nor were they penetrated by boreholes (IGRS-IFP, 1966; BP, 1971).

The beginning of the synrift sequence is represented by the Siniais Limestones and their lateral equivalent, the Louros Limestones (Karakitsios and Tsaila-Monopolis, 1988). Foraminifera, brachiopods, and ammonites in Louros Limestones indicate a Pliensbachian age (Karakitsios, 1990, 1992). These formations correspond to the general subsidence of the Ionian area (formation of Ionian basin), which was followed by the internal synrift differentiation of the Ionian basin marked by smaller paleogeographic units. These paleogeographic units were recorded in the prismatic synsedimentary wedges of the synrift formations and include the Siniais or Louros Limestones, the Ammonitico Rosso or Lower Posidonia beds, the "Limestones with Filaments," and the Upper Posidonia beds (Figure 4). Stratigraphic sections measured throughout the study area display abruptly changing thicknesses in the syntectonic sequences within a few kilometers (Figures 5–8). The opening of the Tethys ocean was accompanied by the formation of a series of north-northwest– and east-southeast–trending conjugate faults. The early Liassic shallow-marine platform was affected by listric block-faulting, which was recorded in the differential subsidence within each small paleogeographic unit (Bernoulli and Renz, 1970; Karakitsios, 1990). The directions of synsedimentary tectonic features (e.g., slumps and synsedimentary faults; Figures 9A, C, D, E, F) indicate that deposition was controlled by structures formed during the extensional tectonic phase. The sedimentation style corresponds, in general, to a half-graben geometry. Prismatic synsedimentary wedges of the synrift formations in the small paleogeographic units (in most cases the units did not exceed 5 km across) vary in thickness east-west. Thus, unconformities are located on top of tilted blocks (Figure 10B), and complete Toarcian to Tithonian successions with Ammonitico Rosso or Lower Posidonia beds at their base are located in the deeper part of the half grabens (Figures 11A, B, C).

It is theoretically possible that the Ionian zone evaporitic substratum halokinesis influenced the synrift mechanism. In fact, in the Pliensbachian, the accumulated thickness of sediments over the evaporitic substratum exceeded 1700 m (Foustapidima Limestones: 200 m; Pantokrator Limestones: 1500 m; Siniais or Louros Limestones: more than 100 m). Under these conditions salt is usually less dense than its overburden so that the unstable condition of a density inversion results. The depth at which the initiation of buoyant rise of the salt through its denser overburden begins is dependent on a number of factors. The presence of lateral heterogeneities such as thickness changes in the overlying sediment or irregularities on the surface of the salt layer are sufficient to trigger upward movement of the low-density salt at relatively shallow depths. Structural heterogeneities may also facilitate the initiation of diapirism at folds or points of structural weakness such as faults. In extensional faulted zones, diapirs tend to form through buoyancy where overburden load is most reduced in the footwall (Jackson and Galloway, 1984; Jackson and Talbot, 1986) (Figure 12). Field data support this mechanism as the most likely driving force behind diapirism in the Ionian basin. The presence of gypsum elements observed in the conglomerate at the base of the Lower Posidonia beds (Toarcian) in the Lithino section (Figure 11E) can easily be explained by evaporitic substratum halokinesis, which led to salt injection along listric faults separating tilted

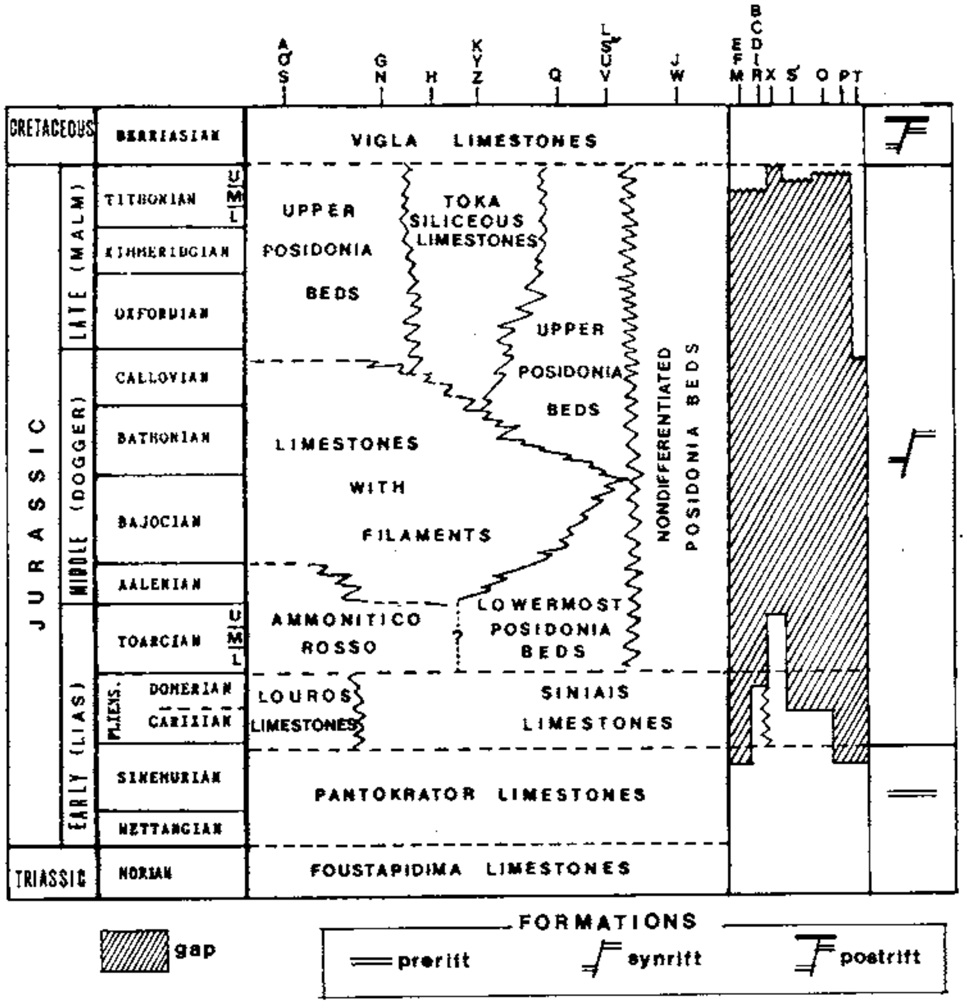


Figure 4—Synthetic diagram showing stratigraphic relationships of the Upper Triassic–Lower Cretaceous formations of the Ionian succession. Letters on top (A through X) correspond to the relative positions of the sections (see Figures 5–8) depending on the succession of synrift formations.

blocks close to the depression where conglomerates were deposited. The Lithino section located in the central Ionian zone is more than 70 km from the adjacent Paxos zone, where the Paxos-1 borehole penetrated anhydrite intercalations in the Liassic dolomites. It is also more than 50 km from the Gavrovo-Tripolitza zone, where sub-Cretaceous formations are unknown and their presence can only be hypothesized. The assumption that gypsum was transported from these zones by turbidity currents for such long distances and was deposited as centimeter-size grains is unreasonable (Karakitsios, 1990, p. 244–245). Consequently, theoretical and field data suggest that the extensional phase provoked halokinesis in the Ionian zone evaporitic substratum (Karakitsios, 1988, 1990, 1992); the halokinesis influenced the synrift mechanism by increasing the extensional fault throws. The combination of these factors resulted in the formation of areas where the evaporitic substratum thickness is maximal (areas with unconformity of the formations deposited during the Toarcian through Tithonian) and areas where the thickness is minimal (areas with Ammonitico Rosso or Lower Posidonia beds). This thickness distribution may considerably facilitate the choice of the favorable locations for boreholes in an attempt to reach, at shallow depths, the unknown subevaporitic Ionian substratum, which may have an oil interest (Nikolaou, 1986).

The stratigraphic relationships of the Upper Triassic to Lower Cretaceous formations of the Ionian succession are given in the synthetic diagram of Figure 4. The geometric characteristics of the beginning of the synrift intrabasinal differentiation into tilted blocks is presented in the paleogeographic and stuctural map of the Toarcian (Figure 13), reconstructed from the lithostratigraphic analysis of the Jurassic formations all over the Ionian zone in the Epirus region (Karakitsios, 1992). In this map, the stratigraphic pinch-outs of the formations

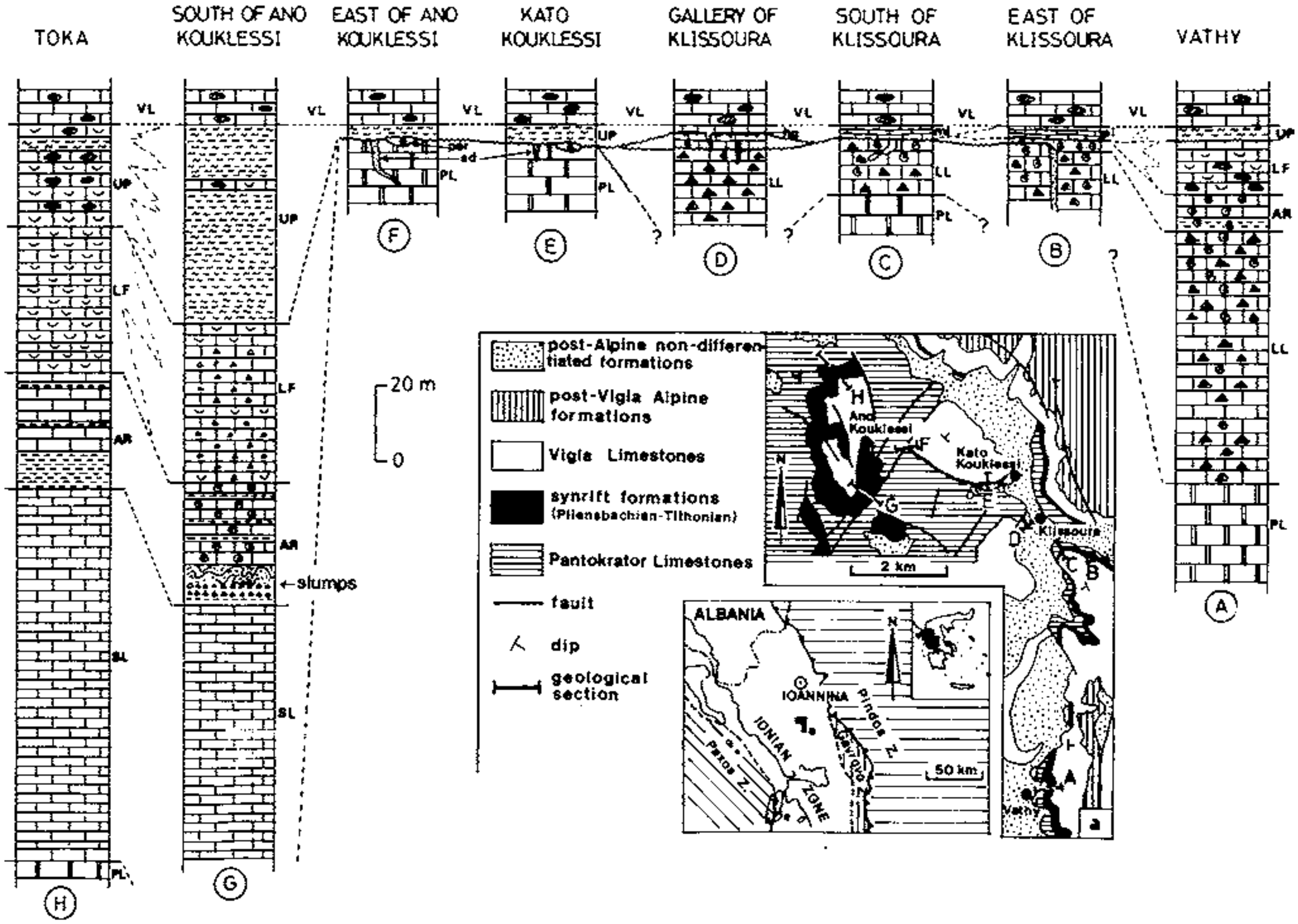


Figure 5—Stratigraphic columns of the central Ionian zone of the Epirus region. VL = Vigla Limestones (base: early Berriasian); UP = Upper Posidonia beds (late Callovian–Tithonian); pl and ml = lateral equivalents of Upper Posidonia beds; hg = hardground; LF = Limestones with Filaments (Bajocian–Callovian); AR = Ammonitico Rosso (Toarcian–Aalenian); par = Ammonitico Rosso (AR) filling the pockets of Pantokrator Limestones (PL); LL = Louros Limestones (Pliensbachian); SL = Siniais Limestones (lateral equivalent of Louros Limestones); PL = Pantokrator Limestones (early Liassic); sd = sedimentary dikes.

deposited from the Toarcian through the Tithonian on each tilted block are parallel to the major faults bordering the tilted block and to the slumps' axes and to minor normal synsedimentary faults observed in the depocenter of the so-formed half grabens. The geometry of the basin explains why the formations deposited from the Toarcian through the Tithonian are organized in prismatic sedimentary wedges. This map contributes significantly to the estimation of the the volume of the Toarcian through Tithonian source rocks ("Lower and Upper Posidonia beds") and consequently to the evaluation of the total petroleum potential of the Ionian basin.

The postrift period was defined by an early Berriasian breakup that is marked by an unconformity at the base of Vigla Limestones (Figure 10B). Sedimentation during the postrift period was synchronous in the whole Ionian basin (Karakitsios, 1990; Karakitsios and Koletti, 1992). The postrift sequence (Vigla Limestones and overlying Alpine formations) largely obscures the synrift structures, and, in some cases, directly overlies the Pantokrator Limestones prerift sequence (Figures 5–7). The deposits of Vigla Limestones do not correspond to a eustatic sea level rise, but to a general sinking of the entire basin (Karakitsios, 1992). The permanence of differential subsidence during the deposition of the Vigla Limestones, shown by the strong variation in thickness of this formation, is probably due to the continuation of halokinetic movements of the Ionian zone evaporitic substratum (IGRS-IFP, 1966; Karakitsios, 1988, 1990, 1991, 1992).

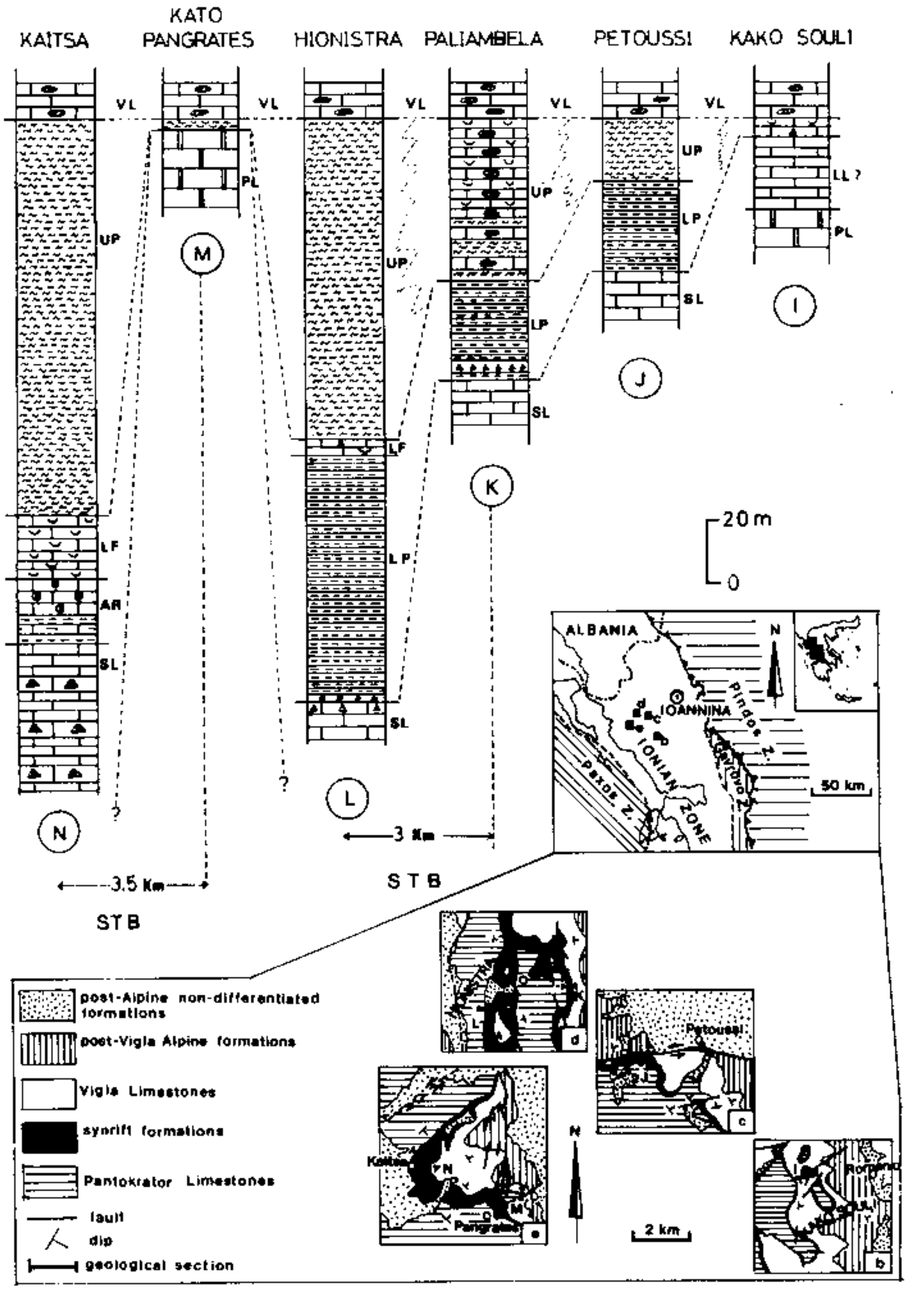


Figure 6—Stratigraphic columns of the central-western Ionian zone of the Epirus region (continuation). For legend, see Figure 5 caption; STB: same tilted block.

ANOXIC SEDIMENTATION IN THE LOWER AND UPPER POSIDONIA BEDS

Two coeval facies exist in the Toarcian of the Ionian zone: an organic-rich black shale known as Lower Posidonia beds (IGRS-IFP, 1966), and the pelagic red nodular limestone with Toarcian ammonites known as Ammonitico Rosso (Renz, 1955). The two facies are confined to the deeper part of generally different half grabens (Figure 13). The Lower Posidonia beds are composed of well-bedded pelagic limestones, marls, and siliceous argillites and show variations in facies and thickness (IGRS-IFP, 1966; Walzebuck, 1982) depending on their position in each half graben (Karakitsios, 1990, 1992). IGRS-IFP (1966) deduced the age of the Toarcian formation from the stratigraphic frame, and recent studies on nannofossil assemblages support that age (Baldanza and Mattioli, 1992).

The total organic carbon of the Lower Posidonia beds measured in different sections of the Ionian zone ranges from 0.45 to 3.64% (Jenkyns, 1988) or from 0.1 to 5% (Baudin and Lachkar, 1990), whereas unpublished data from Public Petroleum

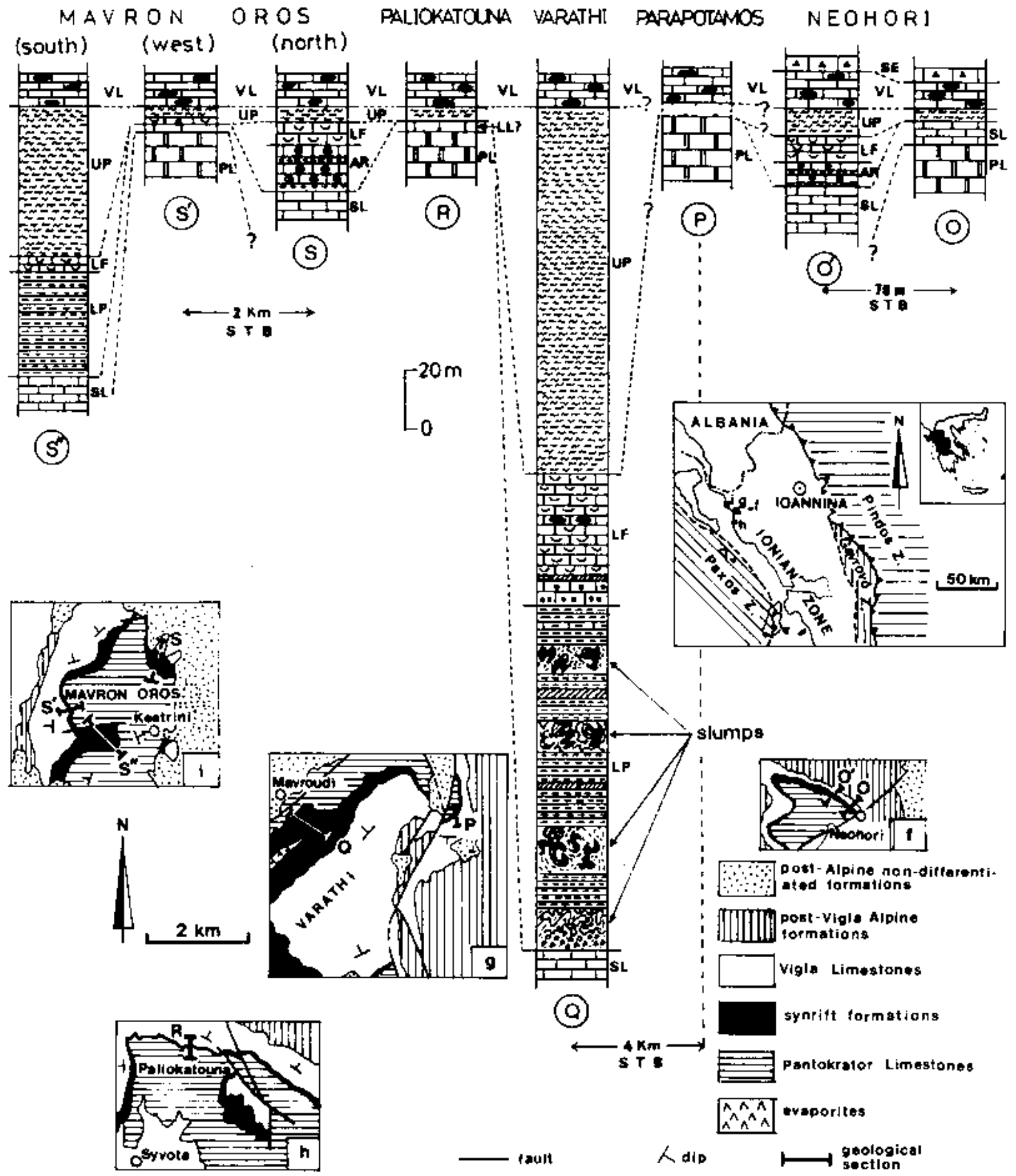


Figure 7—Stratigraphic columns of the western Ionian zone of the Epirus region (continuation). SE: Senonian; VL: Vigla Limestones (base: early Berriasian); UP: Upper Posidonia beds (late Callovian–Tithonian); LF: Limestones with Filaments (Bajocian–Callovian); AR: Ammonitico Rosso (Toarcian–Aalenian); LP: Lower Posidonia beds (lateral equivalent of AR); LL: Louros Limestones (Pliensbachian); SL: Siniais Limestones (lateral equivalent of Louros Limestones); PL: Pantokrator Limestones (Early Liassic); STB: same tilted block.

Corporation (DEP) places the range from 0.1 to 20.05% (N. Rigakis, 1993, personal communication). The accumulation of the organic matter was attributed by Jenkyns (1988) to an early Toarcian anoxic event that was recorded in the different deep-marine Tethyan formations from Austria-Germany, Italy, Greece, Hungary, and Tunisia. Baudin and Lachkar (1990) attributed the organic matter accumulation to a particular type of restricted basin geometry developed during the Toarcian.

The surface occurrences of petroleum in the Ionian zone are generally attributed to Toarcian Lower Posidonia beds source rocks (IGRS-IFP, 1966; Nikolaou, 1986; Baudin and Lachkar, 1990); however, the organic-carbon content in most of the Lower Posidonia beds outcrops is immature (Baudin and Lachkar, 1990). Based on the present geothermal gradient of the Ionian zone (1.5–2.3°C/100 m), which is probably lower than the paleogeothermal gradient, the oil-window depth is estimated to be more than 3500 m. The Upper Posidonia beds (Callovian–Tithonian) (Karakitsios et al., 1988) are generally considered poor in organic matter. However, the Palimbela section (Callovian–Kimmeridgian) (Figure 6) is an organic-rich sequence in which the organic content ranges from 0.5 to 8.6% (Danelian and Baudin, 1990). Additional data are furnished by an exploratory borehole drilled to a total depth of 1530 m in Ioannina (Figure 14). The well was

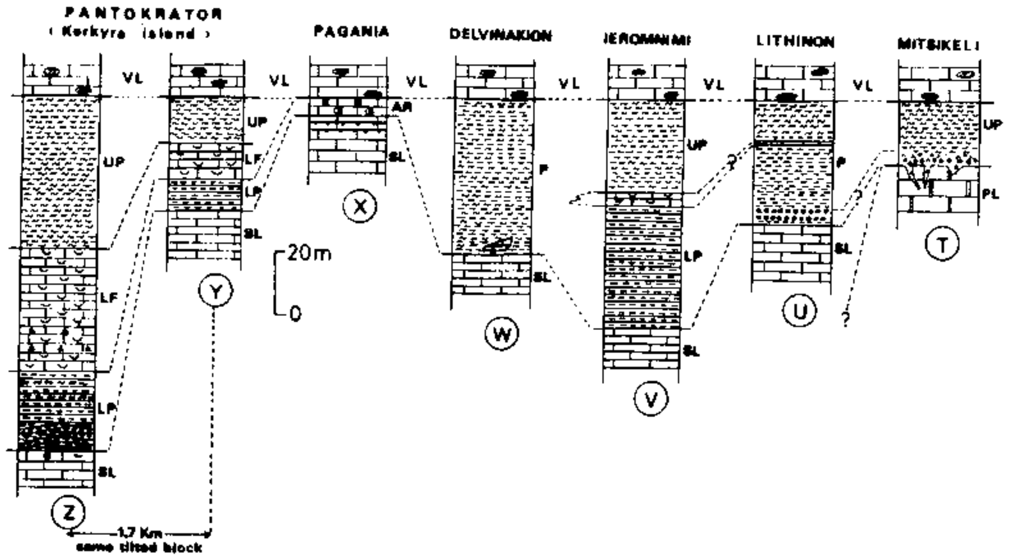
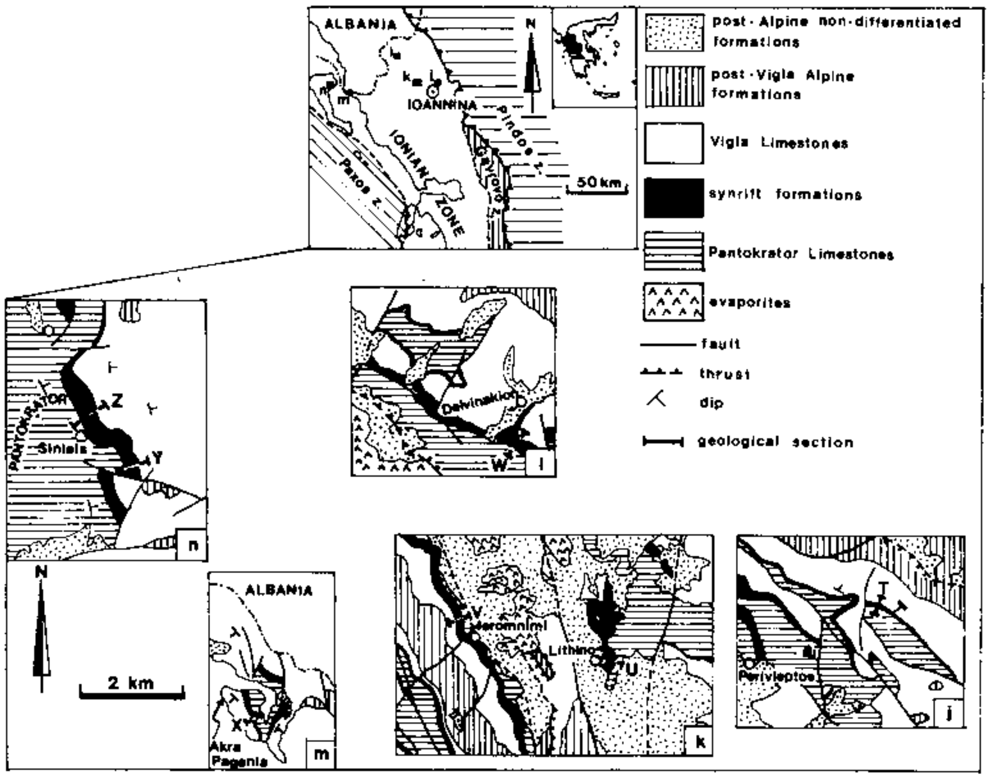


Figure 8—Stratigraphic columns of the northern Ionian zone of the Epirus region (continuation). For legend, see Figure 7 caption; P: nondifferentiated Posidonia beds.



originally drilled to investigate the feasibility of safely injecting treated municipal wastewater of the city of Ioannina at a depth from 1000 to 1530 m. The stratigraphic sequence in the borehole, determined from cuttings and cores, is as follows: from 0 to 300 m Neogene sediments, from 300 to 610 m Vigla Limestones, from 610 to 990 m nondifferentiated Posidonia beds, from 990 to 1150 m Siniais Limestones, from 1150 to 1250 m Pantokrator Limestones, and from 1250 to 1530 m "Triassic breccias." Because the Pantokrator Limestones thickness is generally more than 1000 m thick (IGRS-IFP, 1966; BP, 1971), the contact between Pantokrator Limestones and underlying breccias might be tectonic. The whole nondifferentiated Posidonia beds are rich in organic matter.

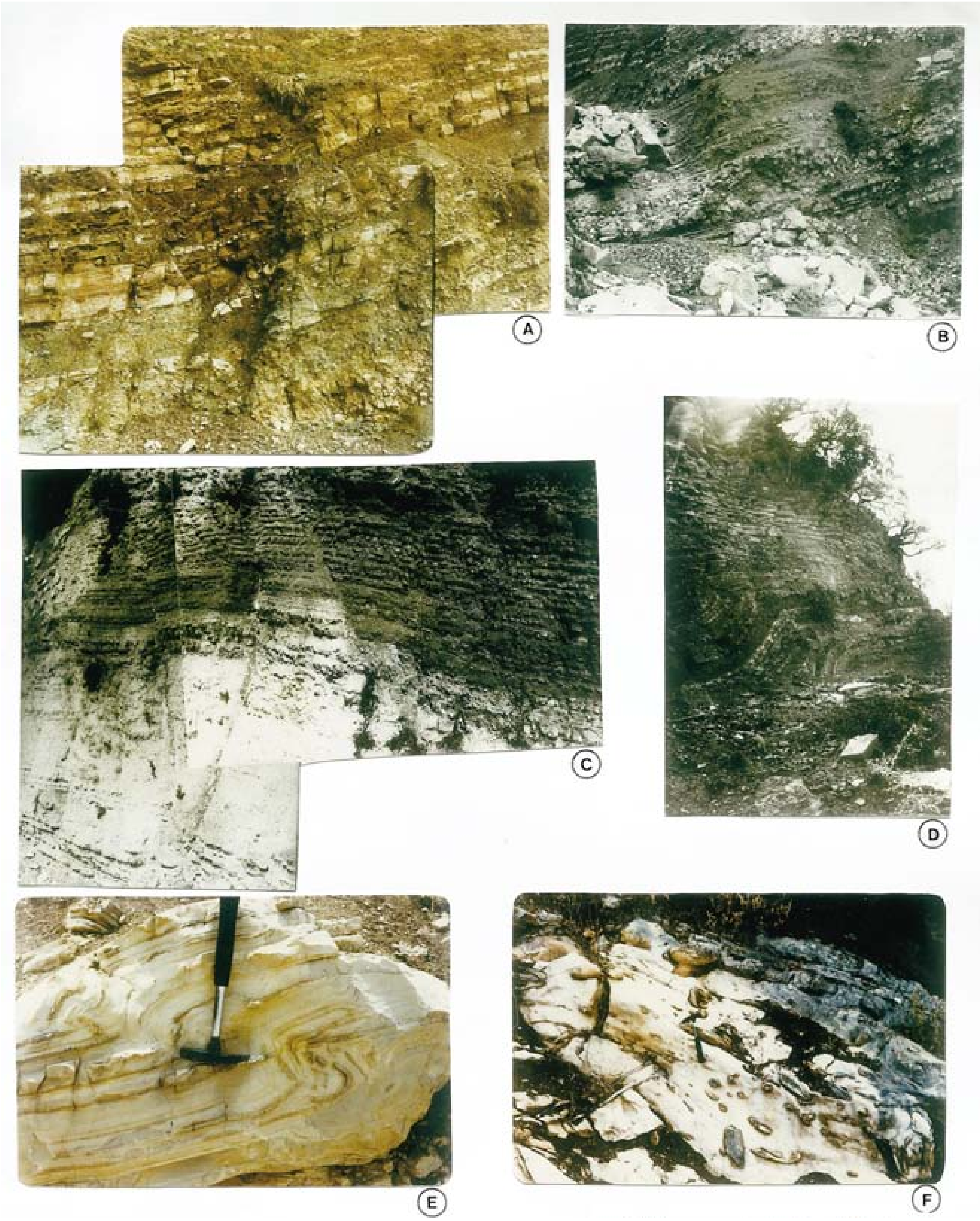


Figure 9—Examples of synsedimentary tectonic features, as seen in outcrop. (A) Synsedimentary normal faults observed in the lower part of Ammonitico Rosso; south of Ano Kouklessi. (B) Breccia zones with big elements (derived from Siniais Limestones) at the base of Ammonitico Rosso; south of Ano Kouklessi. (C) Synsedimentary normal faults observed in the lower part of Ammonitico Rosso; Mikri Vryssi; west of Ano Kouklessi. (D) Slumps observed in the base of Ammonitico Rosso; Mikri Vryssi; west of Ano Kouklessi. (E) and (F) Slumps at the base of the Lower Posidonia beds; Varathi.

The total organic carbon observed in the core sample at the depth of 960 m is 3% for the limestone and 3.8% for the marl (a direct measurement of dip in the alternating limestone and marl of the core sample gave a dip of 45°). This organic carbon appears to be an immature mixture of marine and continental organic matter.

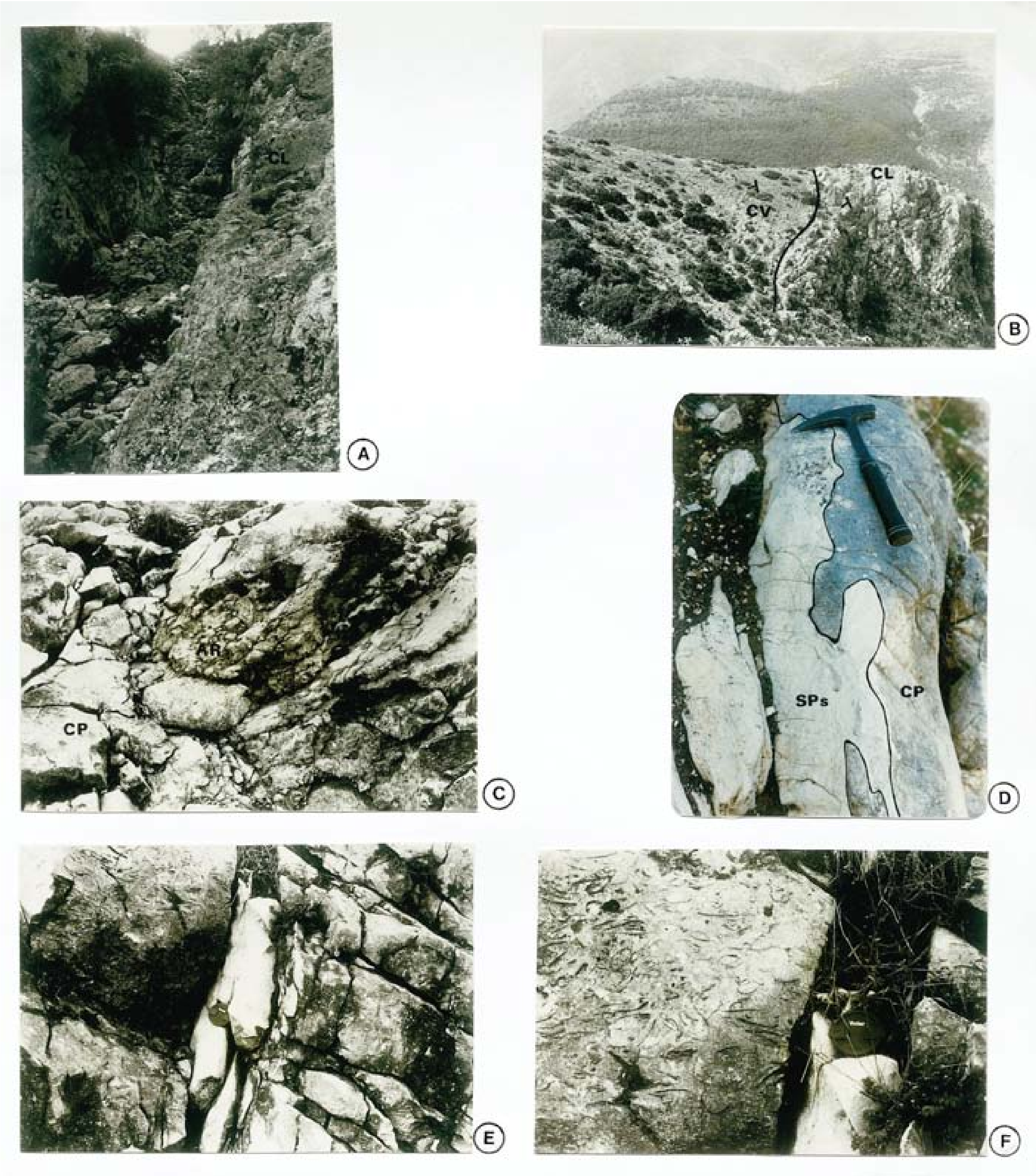
These observations emphasize the local character of anoxic events that occurred in the small subbasins into which the Ionian basin was differentiated during the synrift period. Thus, the anoxic conditions were selectively created during the whole synrift period. The organic matter accumulation in the Ionian zone occurred during the entire Toarcian through Tithonian interval and is directly related to the geometry of the opening of the Ionian basin. The particular type of restricted subbasins that were formed possessed a geometry that favored water stagnation and, consequently, the locally euxinic conditions of the bottom waters. In these areas, the black shale facies (Lower Posidonia beds) was preferentially deposited during the Toarcian. Anoxic conditions were observed locally even during the postrift period in the areas where the Upper Siliceous Zone (Albian–Cenomanian; IGRS-IFP, 1966, and Figure 2) of the Vigla Limestones is well developed (Karakitsios, 1990, p. 186–189). In fact, the total organic carbon of the Upper Siliceous Zone penetrated in the Agios Georgios borehole (central Ionian zone of Epirus region), at a depth from 3150 to 3580 m, ranges from 0.20 to 11.70%, and is mature (N. Rigakis, 1993, personal communication). It is possible that some subbasins were preserved by halokinetic movements even during the postrift period.

ALPINE TECTONICS OF THE IONIAN ZONE

Structural analysis of the Ionian zone shows that orogenesis essentially took place during the major phase of deformation at the end of early Miocene (IGRS-IFP, 1966). The Ionian zone was considered to be the assemblage of anticlines and synclines pushed against each other by a system of moderate westward overthrusts, which were accompanied by a transcurrent fault system (i.e., Petoussi transcurrent fault and Ziros transcurrent fault; Figure 14) that made the thrust mechanism easier (IGRS-IFP, 1966). However, even though eastward-directed thrusts were pointed out by Brunn (1956) and Aubouin (1959), in the internal (eastern) part of Ionian zone, subsequent researchers did not pay attention to these thrust units. Observation of Ionian zone tectonics shows, in fact, that structures (folds and thrusts) in its eastern part transported rocks eastward (Figure 14). This divergence is clearly illustrated in cross sections through Mitsikeli (Figure 15). In the central and western parts of the Ionian zone, structures were displaced westward (Figures 16–18).

In an attempt to evaluate the displacements and examine the manner in which a complex stratigraphy was deformed, balanced cross sections were constructed through an intermediate and a more external part of the basin (Figures 16A and 17A, respectively). The abrupt variations in both the thickness and facies of the Ionian synrift sediments (deposited during the Pliensbachian through Tithonian) together with the evaporitic substratum halokinesis make it difficult to apply simple-type structural analysis because the sedimentary packages do not correspond to layer-cake stratigraphies. Vigla Limestones constitute the first postrift sediments of the basin and consequently were deposited almost horizontally. Thus, for the construction of the restored sections (Figures 16B, 17B), the basal part of the Vigla Limestones has been used as the datum and was restored afterward to horizontal. The balanced cross sections of Figures 16A and 17A illustrate that several times during the compressional phase, faults related to the extensional phase did not reactivate as thrusts, as the classical scheme of inversion tectonics suggests (Figure 19A); but due to evaporitic substratum halokinesis, the most elevated footwalls have been thrust over the preexisting hanging walls (Figure 19B). The amount of shortening (about 20% and 25% in the intermediate and more external part of the basin, respectively) involves a westward growth of horizontal

Figure 10—Examples of bounding surfaces seen in outcrop. (A) Sedimentary dike near top of the Louros Limestones (CL) containing Pliensbachian ammonites. The dike is developed along an open normal synsedimentary fault (fault throw 5 m) (Toarcian–Kimmeridgian); east of Klissoura. (B) The early Berriasian breakup unconformity between the Vigla Limestones (on top) and the Louros Limestones (CL) (which contain Pliensbachian ammonites in their upper part). The contact is marked by 3–4 m of pelagic sediments; south of Klissoura. (C) Ammonitico Rosso (AR) (Toarcian) filling the pockets of Pantokrator Limestones (CP) (early Liassic); east of Ano Kouklessi. (D) Stratigraphic contact between the Upper Posidonia beds (with a thickness reduced up to 3 m) and the Pantokrator Limestones (CP); east of Ano Kouklessi. (E) Sedimentary dike on top of the Pantokrator Limestones composed of calcareous breccias of micritic cement with rare filaments, radiolaria, and spicules of Spongiae. The elements are composed of bioclast lime wackestones with small ammonites, spicules of Spongiae, radiolaria, and rare filaments; east of Ano Kouklessi. (F) Detail of (E).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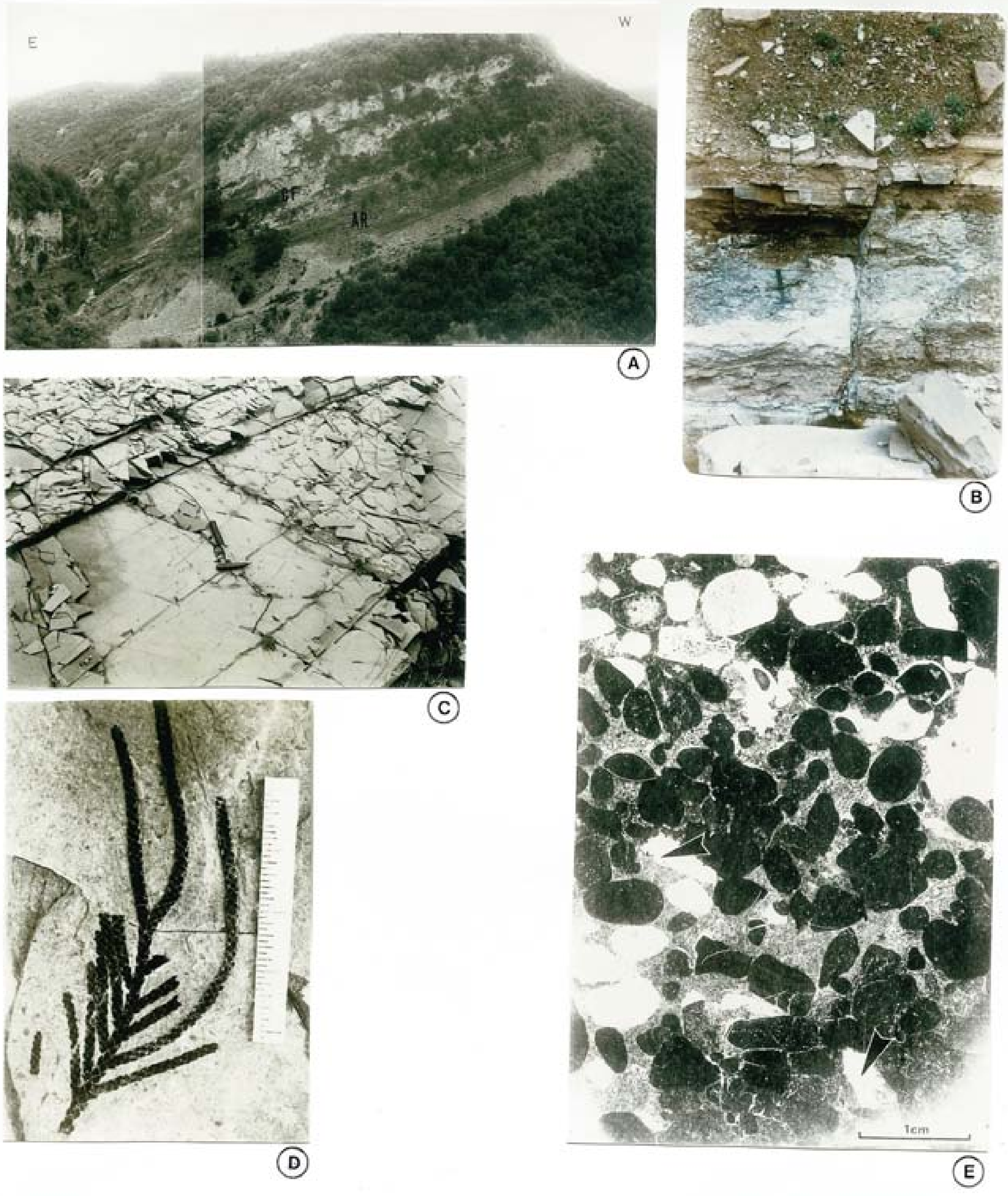

Figure 11—Examples of the different members of the formations deposited during the Toarcian through Tithonian, as seen in outcrop, and thin section from a sample used in this study. (A) Gradual westward reduction in thickness of the formations deposited during the Toarcian through Tithonian (AR: Ammonitico Rosso; LF: Limestones with Filaments); Toka (north of Ano Kouklessi). (B) Lower part of marly Ammonitico Rosso: laminated blue marls (lower part) and red marls (upper part) affected by a synsedimentary fault; Toka (north of Ano Kouklessi). (C) Laminated blue marls (organic-rich black shales) at the base of Lower Posidonia beds; Hionistra, and (D) Coniferous branches (Brachyphyllum nepos saporta) observed in the marls of (C). (E) Thin section from a conglomerate horizon observed in the base of the Lower Posidonia beds; Lithino. This polygenic conglomerate is composed of marly calcareous matrix and three types of elements: the first facies is similar to the Siniais Limestones (lime wackestones with radiolaria), the second facies corresponds to the Louros Limestones facies (micritic limestones with sections of small ammonites and brachiopods), and the third facies is composed uniquely of gypsum.

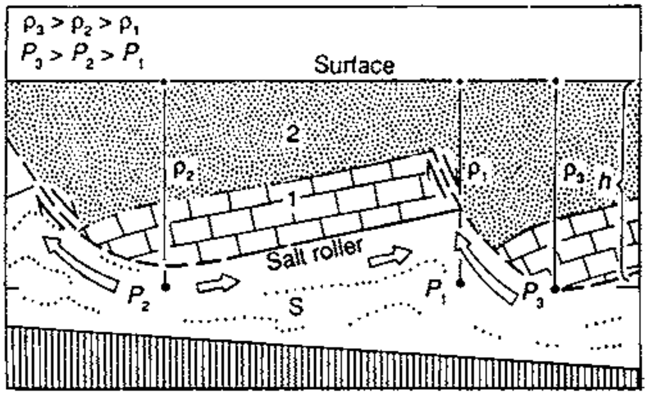

Figure 12—The formation of a salt roller by extensional faulting may trigger the buoyant rise of salt (after Jackson and Galloway, 1984). 1, 2, and 3 are sediment-column densities, P_1, P_2, and P_3 are pressures at three locations, h is the maximum overburden height (in the hanging wall of the fault) on the top of the salt (S). Overburden pressure is relaxed in the footwall (position P1) as a result of the extension.

displacement of the Ionian zone. Since the estimation of the shortening is only based on surface geologogical data, it should be regarded as conservative. However, in all cases the localization of reverse faults and thrusts took place in the location of Jurassic paleofaults. Cross sections mark the coincidence of structural units that were emplaced tectonically 15 Ma with the paleogeographic units that appeared 170 m.y. earlier during the internal differentiation of the Ionian zone. To observe this coincidence, one has to compare the paleogeographic and structural map of the Toarcian (Figure 13) with the map of the present tectonic (Alpine) lines in the Epirus region (Figure 14). The direction of major compression (north-northeast–south-southwest) determined by Cushing (1985) and Karakitsios (1990) diverges from the normal to the direction of thrust surfaces. This divergence suggests that the thrusts, as discontinuities, do not owe their origin to the compressional phase of deformation. It is more likely that these were preexisting discontinuities, which probably originated as listric faults during the extensional phase of the Jurassic. The ancient faults were reactivated in the new compressional state as thrusts with a major horizontal component (Karakitsios, 1990, p. 209–210). The double divergence of Ionian structure (westward in the western and central parts and eastward in the eastern part, respectively) is certainly due to the structures inherited from the extensional Jurassic phase. The extensional Jurassic faults of the external Ionian zone (Apulian side), dipping eastward, were reactivated during the compressional phase as compressional westward displacements, whereas those of the internal Ionian zone (Gavrovo-Tripolitza side), dipping westward, were reactivated as compressional eastward displacements (Figures 13, 14). Listric faults were transformed entirely into transcurrent faults, or reverse faults and/or thrusts, which is consistent with classical inversion tectonics (Figure 19A). Where halokinesis played an important role in the development of the structures, the reverse fault movement occurred only in the upper parts of the faults (Figure 19B). This phenomenon was facilitated by diapiric movements through the tectonic surfaces of the evaporitic base of the Ionian zone. Evidence of a moderate decollement in the subsurface evaporites, in the external domain of the Ionian zone, is found in the Filiates borehole (IGRS-IFP, 1966).

As far as the transcurrent fault system of the Epirus region is concerned, the sinistral transcurrent fault of Petoussi (Figure 14) showed that this system is later than the folds (IGRS-IFP, 1966; Karakitsios, 1990), and the direction of compression related to its activation (northeast-southwest to east-west) is considered to be different from that associated with the thrusts (north-northeast–south-southwest) (Karakitsios, 1990, 1992).

A few seismic lines are available from the Ionian Sea and in the external Ionian zone, and they show a westward growth of horizontal displacement of the Ionian zone, with diapirism expressed more strongly in the same direction (Karakitsios, 1990, p. 255). This is in accordance with the balanced cross sections provided from the field data (see estimated amount of shortening from east to west, Figures 16, 17). Consequently, a moderate decollement in the subsurface evaporites is certain, especially in the external domain of the Ionian zone. Southward, in the Akarnania region, although BP (1971) and Jenkins (1972) considered that the evaporites played a more extensive role in the tectonics than in the Epirus region, they did not deduce a large-scale decollement. However, field data and available seismic lines indicate that it is out of the question to assimilate the decollement deduced here with a major decollement such as that suggested by Guzzetta (1982) in the Epirus region and by Sorel and Cushing (1982) in the Akarnania region.

The tectonic evolution of the Ionian basin drastically affects the probable hydrocarbon traps of the region. In fact, the most suitable reservoir rock lithologies of the Ionian succession are the fractured Eocene limestones and the Triassic, Liassic, and Early Cretaceous dolomites (Nikolaou, 1986). Furthermore, the most appropriate cap-rock lithologies of the Ionian succession are the flysch deposits (Oligocene), the post-Alpine formations (Neogene), and the evaporites (Triassic) (Nikolaou, 1986). The geometry of these deposits (reservoir

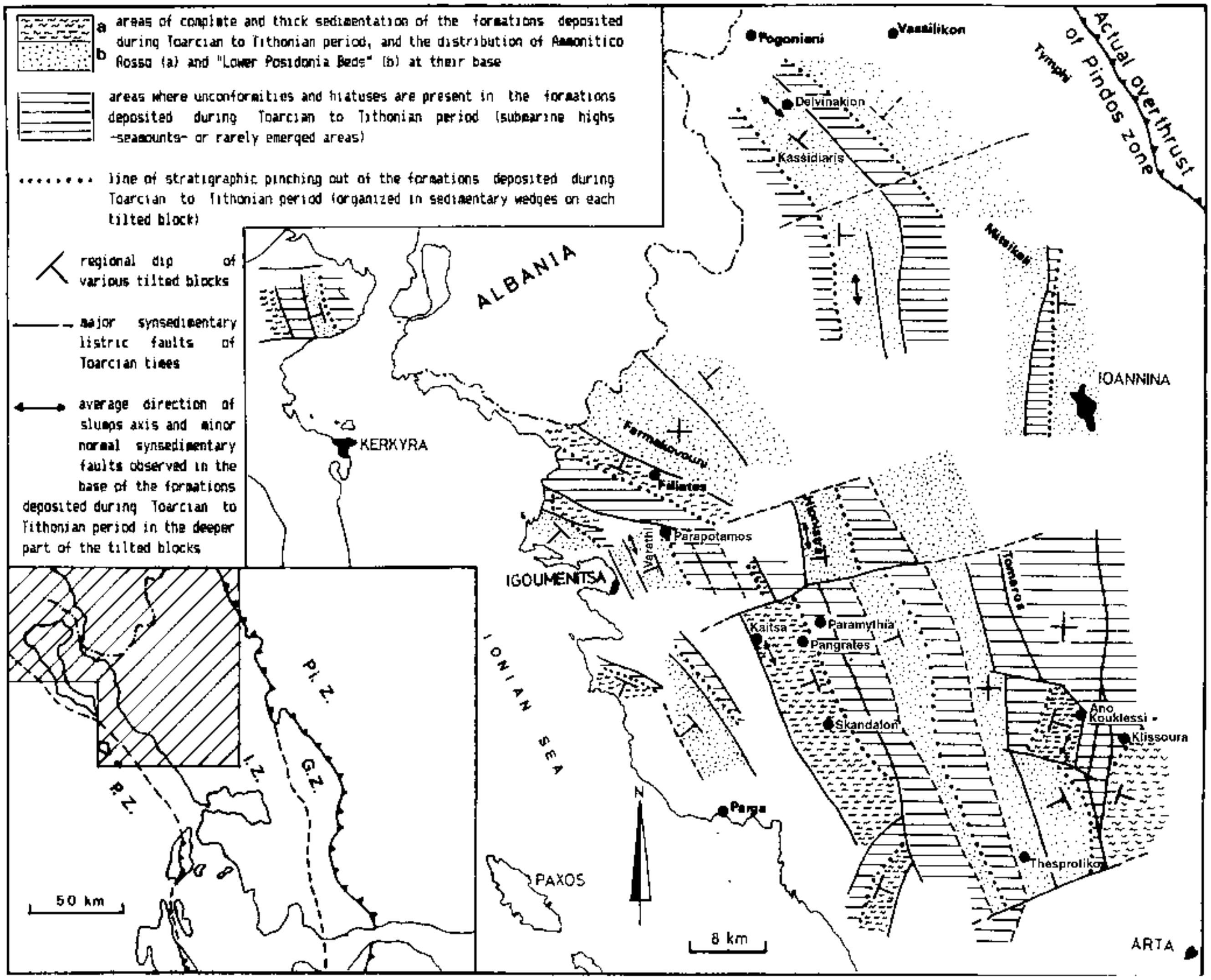

Figure 13—Paleogeographic and structural map of the Toarcian in the Epirus region, northwest Greece (after Karakitsios, 1992).

rocks and cap rocks) is dependent on the previously described inversion tectonics of the Ionian basin. Consequently, this structural style determines the location of the probable hydrocarbon traps in the Ionian basin.

PALEOGEOGRAPHIC AND STRUCTURAL EVOLUTION OF THE IONIAN BASIN IN THE ALPINE CONTEXT

The end of favorable conditions for sulfate precipitation and the beginning of marine sedimentation in the Ionian area was marked by deposition of the Foustapidima Limestones during the Late Triassic (Karakitsios, 1990). During the early Liassic, a huge carbonate platform bordering the southern Tethys ocean was established over the whole of western Greece. This platform was characterized by strong subsidence which was balanced by intensive carbonate sedimentation in a very shallow water sedimentary environment (Bernoulli and Renz, 1970; Karakitsios, 1990). The resulting accumulated carbonate succession, the Pantokrator Limestones, is more than 1000 m thick (IGRS-IFP, 1966; BP, 1971).

During the early Pliensbachian, the initial shallow carbonate platform began to break up (Karakitsios, 1990). The first general deepening of the Ionian area was recorded by the Siniais and the Louros Limestones deposits. The extension that involved this deepening was probably expressed by border

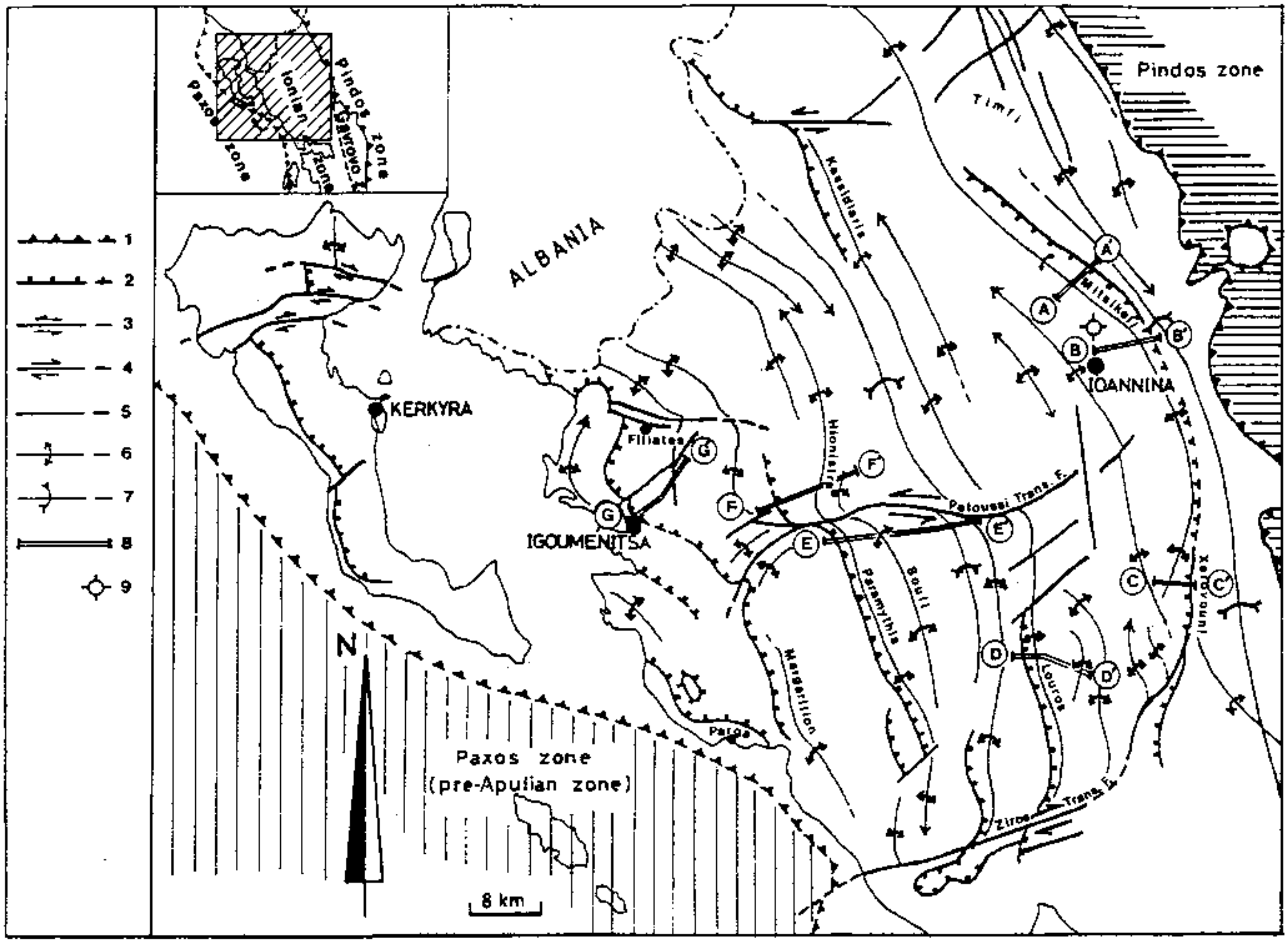

Figure 14—Structural map of the Epirus region (after IGRS-IFP 1966, modified): 1 = Pindos overthrust, 2 = thrust or reverse fault, 3 = sinistral transcurrent fault, 4 = dextral transcurrent fault, 5 = fault, 6 = anticlinal axis, 7 = synclinal axis, 8 = locations of structural cross sections, 9 = Ioannina exploratory borehole.

faults that separated the Ionian basin from the adjacent Paxos (in the west) and Gavrovo-Tripolitza (in the east) zones where the production of platform carbonates persisted throughout the Jurassic. Louros Limestones and their lateral equivalent, the Siniais Limestones, correspond to the first synrift sediments of the Ionian series. The Siniais facies occupied the axial part of the Ionian basin whereas the Louros facies occupied its bordering areas.

From the end of the Pliensbachian to the end of the Toarcian, the continuation of extension was accompanied by intense block faulting that led to the internal differentiation of the Ionian basin. Listric faults associated with this phase caused the separation of the initial basin into a number of small (2–10 km across) paleogeographic units that were subjected to differential subsidence. These paleogeographic units were characterized by tilted blocks that formed half grabens. In the deeper part of the half grabens, Ammonitico Rosso (Figure 11A, B) or Lower Posidonia beds (Figure 11C) were deposited. These deposits were accompanied by products of submarine or subaerial erosion derived from the top of the same tilted block as well as from the tops of the adjacent tilted blocks. The deposits derived from the tops of adjacent tilted blocks consisted of breccia with big elements or big blocks detached from the fault scarps that fell into the depressed part of the tilted block (Figure 9B). On the top of the tilted blocks, sedimentary dikes (Figure 10A, E, F), hardgrounds, and evidence for sedimentary hiatuses are found (Figure 10B, C, D). These tops constitute either a submarine high or, in places, emergent relief. The case of emergent relief is attested to by the presence of coniferous branches (Figure 11D) observed in the base of Lower Posidonia beds (e.g., Hionistra section; Karakitsios, 1990, p. 122, 125). The observed

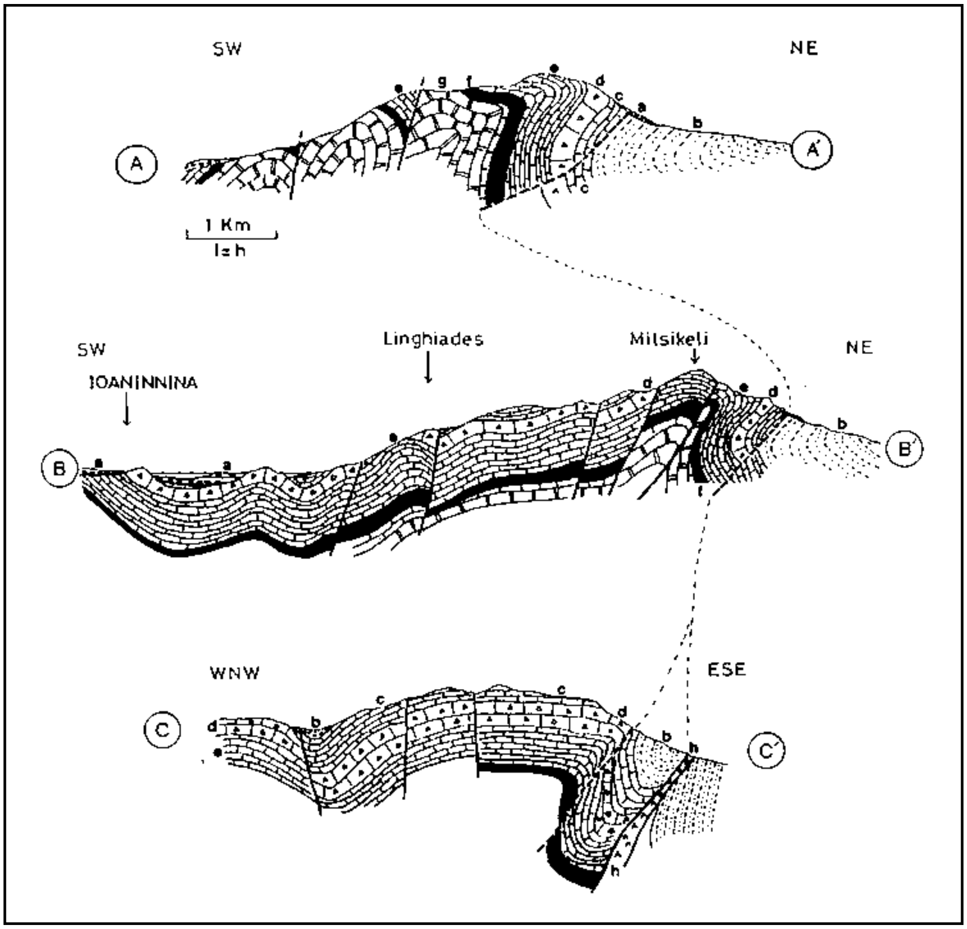

Figure 15—Sections of Mitsikeli and its southern prolongation (A and B) Mitsikeli anticline. (C) Arakhthos anticline. For locations of sections see Figure 14 (based on field data and geological maps of IGRS-IFP, 1966). a = Pliocene–Quaternary, b = flysch, c = Paleocene–Eocene, d = Upper Senonian, e = Vigla Limestones, f = Posidonia beds, g = Pantokrator Limestones, h = evaporites.

coniferous branches involve terrestrial drifts that can be explained only if the area of Lower Posidonia beds sedimentation was close to an emergent area favorable to coniferous life. Gypsum injections into the fault surfaces were possible if fault throws were strong (i.e., Lithino section and Figure 11E). The same geometry of the basin persisted, with minor modifications, until the Late Jurassic with the conditions of sedimentation becoming more calm with the progressive filling of the depressed parts by the formations deposited during the Toarcian and Tithonian.

During the early Berriasian, a general sinking of the entire basin is attested to by the onset of the deposition of pelagic Vigla Limestones in the entire Ionian zone. Apart from halokinetic movements, which probably caused the variation in thickness of Vigla Limestones, the pelagic conditions persisted (pelagic sedimentation continued to accompany clastic deposition derived from the adjacent Gavrovo-Tripolitza and Apulian platforms) until the late Eocene, when flysch sedimentation began.

At the end of the early Miocene, the major compressional phase that affected the Ionian zone reactivated the preexisting Jurassic extensional fault system with a reversed sense of motion. Nevertheless, because of the halokinesis, this reactivation did not always follow the classical scheme of inversion tectonics. Listric faults were transformed into reverse faults, thrusts, or transcurrent faults. This phenomenon was facilitated by diapiric movements through the tectonic surfaces of the evaporitic base of the Ionian zone. A moderate decollement in the subsurface evaporites, especially in the external domain of the Ionian zone, is certain. However, field data and available seismic lines exclude a major decollement. The symmetry of the Ionian basin associated with the extensional phase of Jurassic times is manifest in the double divergence of its compressional structure (westward in the west and eastward in the east).

CONCLUSIONS

The Ionian zone is a good example of inversion tectonics in a basin with evaporitic substratum. Its paleogeographic and structural evolution is

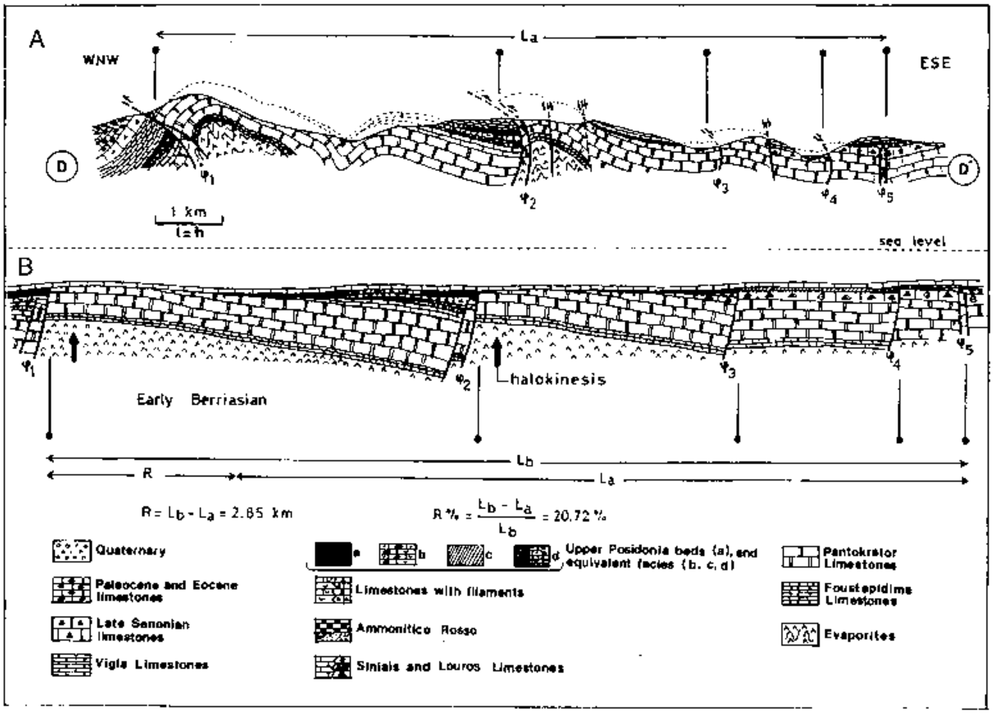

Figure 16—Cross section DD′ of Derviziana-Klissoura: (A) balanced and (B) restored. See Figure 14 for location.

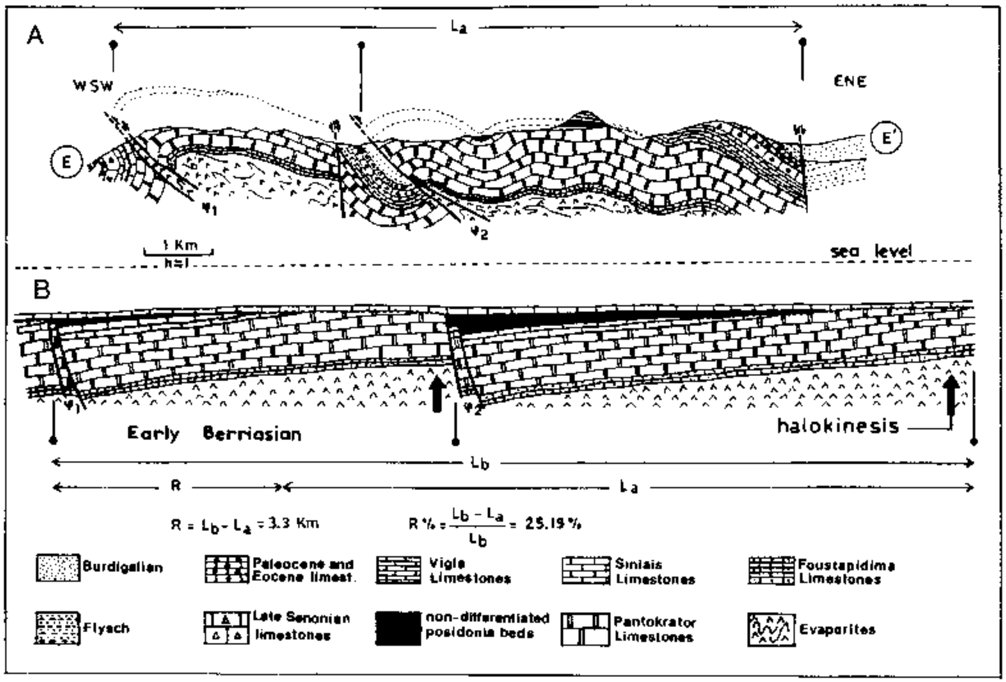

Figure 17—Cross section EE′ of Paramythia: (A) balanced and (B) restored. See Figure 14 for location.

sufficiently comparable to the Umbria-Marche zone of the North Appenines (Alvarez, 1989; Barchi et al., 1989; Cecca et al., 1990).

Halokinesis of the evaporitic substratum of the Ionian basin has had a considerable influence on the synrift mechanism since the Toarcian. Two important results of this influence are that (1) several times during the compressional phase, faults related to the extensional phase did not reactivate as thrusts, as the classical scheme of inversion tectonics suggests, but the most elevated footwalls were thrust over the preexisting hanging walls, and

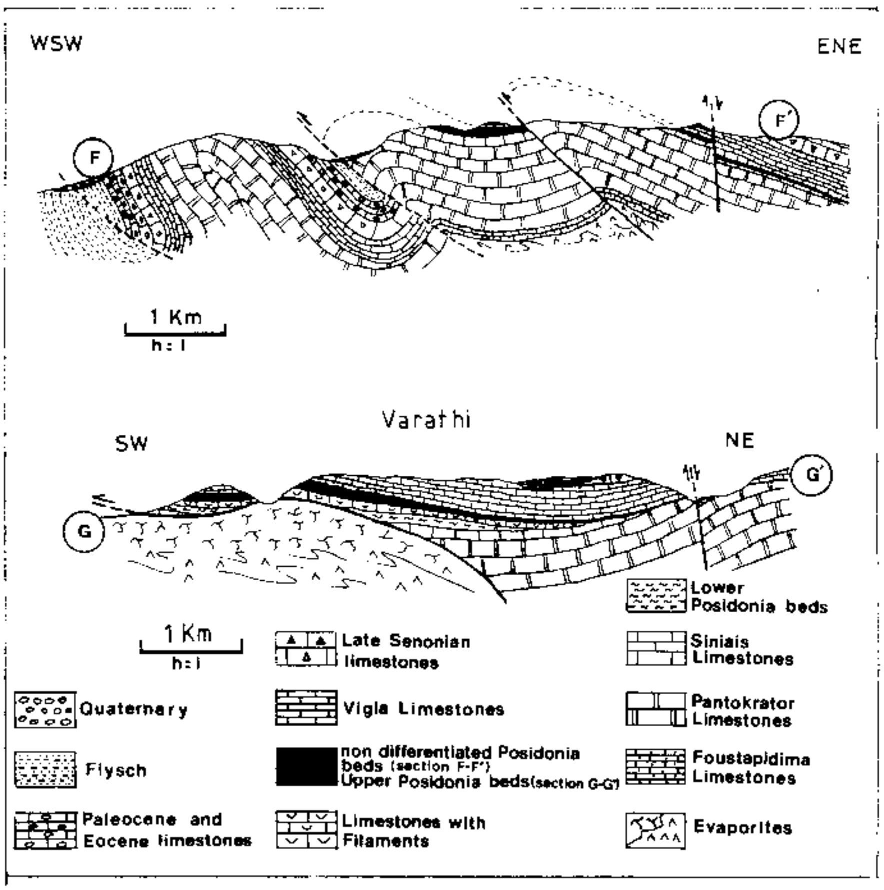

Figure 18—Cross sections FF′ of Zoumpani-Hionistra and GG′ of Varathi. See Figure 14 for location.

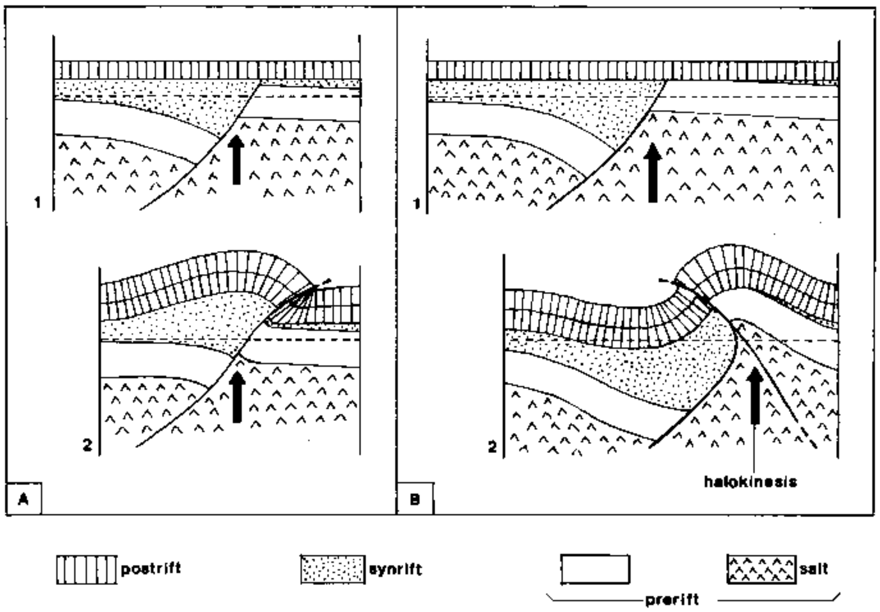

Figure 19—Examples of inversion tectonics affecting a half-graben system with evaporitic substratum (Ionian zone, northwest Greece). (A) Classical inversion tectonics. (B) Particular case of inversion tectonics observed at locations where the evaporitic substratum halokinesis was more expressed and consequently the footwalls of the extensional phase were above average. Therefore, during the compressional phase these most elevated footwalls have been thrust over the preexisting hanging walls. (A1) and (B1) correspond to the beginning of the postrift period; (A2) and (B2) correspond to the end of postrift deposition and show the subsequent inversion geometries.

(2) the locations of maximal evaporitic substratum thickness are situated below the areas with unconformities of the formations deposited during Toarcian to Tithonian, whereas the locations of minimal evaporitic substratum thickness are encountered below the areas where Ammonitico Rosso or

Lower Posidonia beds are well developed. The latter may considerably facilitate the location of boreholes in an attempt to reach, at lesser depths, the unknown subevaporitic Ionian substratum, which may be of oil interest.

The accumulation of organic matter in the Lower and Upper Posidonia beds of the Ionian zone during the Toarcian and Tithonian is directly related to the geometry of the synrift period of the Ionian basin. The geometry of the restricted subbasins that formed favored water stagnation and, consequently, the development of locally euxinic conditions of the bottom waters. Anoxic conditions occurred locally even during the postrift period in the areas where the Upper Siliceous Zone (Albian–Cenomanian) of the Vigla Limestones formation is well developed; these areas probably represent subbasins that were preserved by the continuation of halokinetic movements during the postrift period.

The opening and the inversion tectonics of the Ionian basin influenced both the source rocks and the probable hydrocarbon traps of the Ionian zone.

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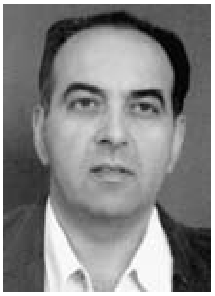
ABOUT THE AUTHOR

Vassilis Karakitsios



Vassilis Karakitsios, born in 1952, is associate professor in the Department of Geology, National University of Athens. His professional experience includes stratigraphy and tectonics of continental collision zones of the eastern Mediterranean and basin analysis of the external Hellenic zones. His academic background includes a Diploma in Geology (1976) from the National University of Athens, a Diplôme d'Etudes Approfondies (1977) and a Diplôme de Docteur de 3e cycle (1979) in structural geology, and a Diplôme de Docteur (1990) in earth sciences from the Pierre et Marie Curie (Paris 6) University.