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Tectonostratigraphic Evolution of the North Anatolian Palaeorift (NAPR): Hettangian–Aptian Passive Continental Margin of the Northern Neo-Tethys, Turkey

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Abstract: The Jurassic-Early Cretaceous stratigraphical, sedimentological and structural evolutionary history of the northern branch of Neo-Tethys is recorded and preserved within an average 6-km-thick basin fill. The basin fill is well–exposed as discontinuous inliers of varying size and thickness within, and to the north of, the Late Cretaceous–Early Tertiary İzmir-Ankara-Erzincan Suture throughout northern Turkey.

This basin fill unconformably overlies a substratum comprising the deformed rock assemblages of the Triassic Karakaya palaeorift and its Hercynian basement. Detailed studies based on geological mapping at a 1/25,000 scale and stratigraphic sections measured at 14 type localities show that this infill includes stratigraphical and sedimentological key units, which comprise a 0.1–3.8-km-thick basal association (proto- and syn-rift infill), a 0.1–1.5-km-thick platform association, a 1.6-km-thick basinal association (transitional to post-rift stage infill), and syn-depositional megabreccias and slumps. This basin fill is shaped by a series of structural units which result from four distinct phases of extensional faulting. These structural units comprise interior grabens, subplatforms, platforms, palaeohighs, rift-related unconformities (proto-, syn- and post-rift unconformities) and rift-related faults. The latter include older growth faults, inner and outer master fault systems, and a marginal fault complex.

This thick Jurassic-Lower Cretaceous basin fill has been interpreted to be the product of a palaeorift, designated here as the north Anatolian palaeorift (NAPR). The young substratum and the key stratigraphical and structural elements support the interpretation that the NAPR comprised the south-facing passive continental margin of north Neo-Tethys in northern Turkey. It had the character of a supradetachment basin and can be explained by the evolution of a three-stage graben model during the Hettangian-Aptian time interval.

Key Words: tectonostratigraphy, palaeorift, three-stage graben model, supradetachment basin, Hettangian–Aptian, northern Neo-Tethys, Northern Turkey

Kuzey Anadolu Paleoriftinin (KAPR) Tektonostratigrafik Evrimi: Kuzey Neo-Tetiz'in Hettanjiyen–Apsiyen Pasif Kıta Kenarı, Türkiye

Özet: Neo-Tetiz'in kuzey kolunun Jura–Erken Kretase'deki stratigrafik, sedimentolojik ve yapısal evrimi 6 km kalınlığındaki bir havza dolgusu tarafından kayıd edilmiş ve korunmuştur. Bu istif, kuzey Türkiye boyunca yer alan Geç Kretase–Erken Tersiyer yaşlı İzmir-Ankara-Erzincan kenedi içinde ve onun kuzeyinde, değişik boyut ve kalınlıklı süreksiz aşınım pencereleri olarak yüzeyler.

Jura–Kretase yaşlı havza dolgusu daha yaşlı bir temel üzerinde uyumsuz olarak yer alır. Bu yaşlı temel, Triyas yaşlı Karakaya paleoriftinin deformasyon geçirmiş kaya topluluğundan ve riftin Hersiniyen yaşlı temelinden oluşur. 1/25,000 ölçekli jeolojik haritalama ve 14 tip alanda ölçülen stratigrafik kesitlere dayalı ayrıntılı çalışmalar, havza dolgusunun, önemli bazı stratigrafik ve sedimentolojik anahtar birimler içerdiğini göstermiştir. Bu anahtar birimler başlıca 0.1–3.8 km kalınlıkta taban topluluğu (rift öncesi-riftle yaşıt dolgu), 0.1–1.5 km kalınlıkta platform topluluğu, 1.6 km kalınlıkta havza topluluğu (geçiş evresi-rift sonrası dolgu), depolanma ile yaşıt megabreşler ve oturma-yıkılma yapılarından oluşmaktadır. Havza dolgusu, dört ayrı normal faylanma evresinde gelişmiş bir seri yapısal birim tarafından şekillenmiştir. Bu yapısal birimler başlıca iç graben, ikincil platform, platform, paleoyükselti, rift uyumsuzluğu (rift öncesi, riftle yaşıt ve rift sonrası uyumsuzluklar) ile rift faylarından oluşur. Rift fayları büyüme fayları, iç ve dış graben ana fay sistemleri ve kenar fay komleksleriyle temsil edilir.

6 km kalınlıktaki Jura–Erken Kretase yaşlı bu havza dolgusu, Kuzey Anadolu paleorifti (KAPR) olarak adlanan bir paleoriftin ürünü olarak yorumlanmıştır. Gerek havza temeli gerekse havza içinde gelişmiş stratigrafik ve yapısal birimler, Kuzey Anadolu paleoriftinin, kuzey Türkiye'de, Neo-Tetiz'in güneye bakan pasif kıta kenarını oluşturduğu yorumunu desteklemektedir. KAPR, sıyrılma fayı üzerinde gelişmiş bir havza karakterine sahip olup, bu havzanın evrimi, Hettanjiyen–Apsiyen sırasında etkin olmuş üç evreli bir graben modeli ile açıklanabilir.

Anahtar Sözcükler: tektonostratigrafi, paleorift, üç evreli graben modeli, sıyrılma fayı üzeri havza, Hettanjiyen–Apsiyen, Kuzey Neo-Tetiz, kuzey Türkiye

Introduction

The past decade has produced a revolution in basin analysis. This revolution resulted from both the recognition of highly extended terrains, and the development and refinement of some analytical techniques such as numerical modelling, isotope stratigraphy, improved geochronology, magneto-stratigraphy and multi-channel seismic data. These advances have facilitated a distinction between traditional rift basins and supradetachment basins (Friedmann & Burbank 1995).

The African rift system, Gulf of Suez, northern Red Sea, the Newark rift system and Lake Baikal are all examples of classical rift basins (Hempton 1987; Morley 1989; Leeder 1995). They developed within intracontinental terrains such as the main continental bulk of the African plate and remote from young orogens. Therefore, they are associated with normal-thickness, cold and old lithosphere, and are characterized by alkalic to tholeiitic magmatism, steeply dipping bounding faults, and low but long-term extension (Morley 1989; Leeder 1995). In contrast, supradetachment basins develop over young orogens and/or back-arc settings, and occur in highly extended terrains such as the United States Cordillera, the Cyclades islands in the northern Aegean Sea, and the present-day Western Anatolian region in Turkey. Therefore, they are associated with thick or overthickened, hot and young crust, and are characterized by tholeiitic to calc-alkalic magmatism, high but relatively short-term extension and low-angle bounding faults (Friedmann & Burbank 1995).

One well-preserved example of a supradetachment basin is the Jurassic–Early Cretaceous north Anatolian palaeorift (NAPR). It is exposed as discontinuous inliers of varying size within, and to the north of the Late Cretaceous–Early Tertiary İzmir-Ankara-Erzincan suture (İAEZ), throughout northern Turkey (Figure 1A). The NAPR comprises the northern passive continental margin of the Northern Neo-Tethys, which began to develop by rifting during the Hettangian in the back-arc setting of the southerly subducting Palaeo-Tethys (Şengör 1979). Sedimentological evidence for the opening history of the northern branch of Neo-Tethys and its Late Jurassic–Aptian evolution in north-western Anatolia have been previously documented by Görür *et al.* (1983) and Koçyiğit *et al.* (1991a).

In the present paper we discuss and summarize the key structural elements and tectonostratigraphic components of the NAPR based mainly on field data obtained throughout northern Turkey from Halılar in the west to Yusufeli in the east (Figure 1). In addition, special attention is focused on the tectonostratigraphic characteristics of the intra-carbonate platform palaeohighs. Detailed 1/25,000 scale geologic maps and the measured sections showing the different lithofacies, synsedimentary structural elements, such as growth faults, slumps, slope breccias, olistostromes; grabens and palaeohighs were documented and published separately in our previous studies carried out at various localities through the northern Turkey (Koçyiğit 1987; Koçyiğit & Altıner 1990; Koçyiğit *et al.* 1991a, b; Altıner *et al.* 1991). In order to avoid repetition of details of the local data, only the simplified pattern of measured stratigraphical sections (1–14 in Figure 1) will be used in the present paper. For detailed information the readers are referred to the afore-mentioned papers. One exception to this is the İneköy study that includes new findings on Jurassic volcanic rocks of the Ankara region. For this reason, a detailed stratigraphic section of the İneköy (Ankara) area will also be given here.

Summary of Late Palaeozoic–Early Mesozoic Palaeogeographic Evolution and Tectonic Setting of the NAPR

Although views on Palaeo-Tethys, Neo-Tethys and their subduction directions remain controversial, all of these

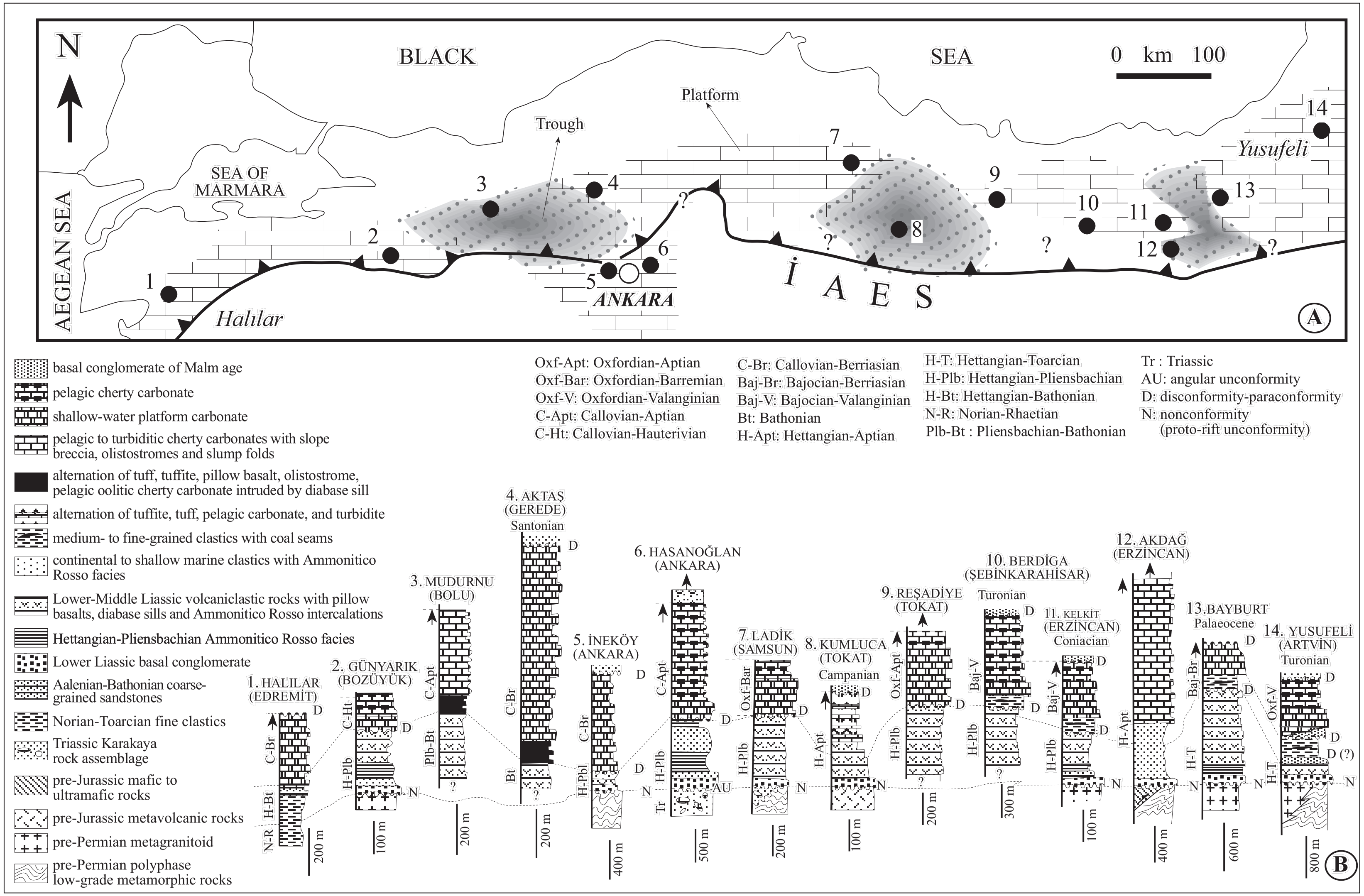

Figure 1. **(A)** Simplified map showing type localities (Numbers 1 through 14) of the North Anatolian Palaeorift infill. İAES- İzmir-Ankara-Erzincan Suture; **(B)** Stratigraphical sections showing both the lateral and vertical facies distribution of the North Anatolian Palaeorift infill. Sections 7, 9, 10, 11, 13 and 14 were compiled in addition to authors own field studies and measurements, from Öztürk (1973), Seymen (1975), Rojay (1985), Gürsoy (1989), Wedding (1963), and Baydar *et al.* (1967), respectively.

ideas can be categorized into two general groups: (1) Those studies referring to Tethys as a single oceanic basin system subducting northward in the eastern Mediterranean region that persisted at least from Permian until Early Tertiary times (Robertson & Dixon 1984; Robertson *et al.* 1991; Stampfli *et al.* 1991; Dercourt *et al.* 1993; Ustaömer & Robertson 1997; Okay & Tüysüz 1999); (2) According to other groups of studies, there was a super oceanic basin system composed of a southward subducting Palaeo-Tethys of Carboniferous-Mid Jurassic age and a northward subducting Neo-Tethys of Triassic-Early Tertiary age (Şengör 1979; Şengör & Yılmaz 1981; Koçyiğit 1987, 1991a; Koçyiğit *et al.* 1991a; Yılmaz *et al.* 1997). Our recent field, stratigraphical, palaeontological and sedimentological data, which will be presented in detail below, support the second general view.

During the Carboniferous–Permian interval, the northern part of present-day Turkey (Pontides), with the exception of the İstanbul-Zonguldak Zone and the Hercynian basement of the future Sakarya Continent, constituted the northern convergent margin of Gondwanaland (Robertson & Dixon 1984; Şengör *et al.* 1984; Yılmaz *et al.* 1997; Altıner *et al.* 2000). However, during the Early and Mid-Triassic, the Hercynian basement of the future Sakarya Continent included a magmatic arc situated on the southerly subducting ocean floor of Palaeo-Tethys and became the site of sedimentation for the Karakaya marginal basin (Şengör 1979; Şengör & Yılmaz 1981; Yılmaz *et al.* 1997). During the Late Triassic, the Karakaya rock assemblage and its Hercynian basement were intensely deformed and became the northern part of Gondwanaland owing to the closure of the Karakaya marginal basin (Figure 2A). In Hettangian time, a new rifting process began within this very young and flexurally to isostatically-uplifted orogen (Figure 2B,C) (Görür *et al.* 1983; Koçyiğit 1991a; Altıner *et al.* 1991; Koçyiğit et al. 1991a; Yılmaz *et al.* 1997). This new rifting (Figure 2D) triggered the development of both the NAPR and northern Neo-Tethys (İzmir-Ankara-Erzincan Ocean). The former is here interpreted as a large and composite supradetachment basin constituting the northern passive continental margin of Northern Neo-Tethys in Jurassic–Early Cretaceous times (Figure 2E).

Stratigraphic Outline

Stratigraphical and sedimentological characteristics of the NAPR infill were studied in detail by both geological mapping at a 1/25,000 scale and by stratigraphical sections measured at 14 type localities (Figure 1). The basement rocks are beyond the scope of this paper, and only brief information about them will be given below.

Basement Rocks

At locations 3, 4, 9 and 10 in Figure 1, the basement of the north Anatolian palaeorift is not observed. In contrast, at the remaining type localities, the basement consists of one or more of the following rocks of the Palaeo-Tethyan orogen: (a) pre-Permian (?) polyphase low-grade metamorphic rocks, (b) pre-Permian metagranitoid, (c) pre-Jurassic metavolcanic rocks, (d) pre-Jurassic ultramafic rocks to mafic rocks, and (e) the Triassic Karakaya palaeorift fill (Figure 1B).

(a) The first group of widespread basement rocks comprises polyphase low-grade metamorphic rocks. They consist mainly of green schists, marble, metavolcanic rocks, metamafic to ultramafic rocks, and rarely higher-grade metamorphic rocks such as gneiss, amphibolite and glaucophane schist. They are well observed at locations 5, 7, 8, 12 and 14 in Figure 1. Their age is still controversial (Okay 1983; Koçyiğit *et al.* 1991b; Genç & Yılmaz 1995).

(b) The Hercynian granitoids are widely exposed in the west and east of northern Turkey as shown in locations 2, 11, 13 and 14 in Figure 1. They are represented mainly by metagranite, leucogranite and granodiorite cut by aplite and microdiorite dikes. In places, the granitoids intruded the low-grade metamorphic rocks. According to disputable radiometric dating carried out by Çoğulu (1975) and Bergougnan (1987), their ages may range from 280 My to 360±2 My (Carboniferous).

(c) The pre-Jurassic metavolcanic rocks are exposed only in location 8 (Figure 1). They are mostly andesitic in composition.

(d) The fourth basement rock suite is composed of pre-Jurassic ultramafics and mafic rocks. They are observed at location 12 (Figure 1) where they are in thrust-fault contact with low-grade metamorphic basement. These rocks occur as tectonic slabs or sheets

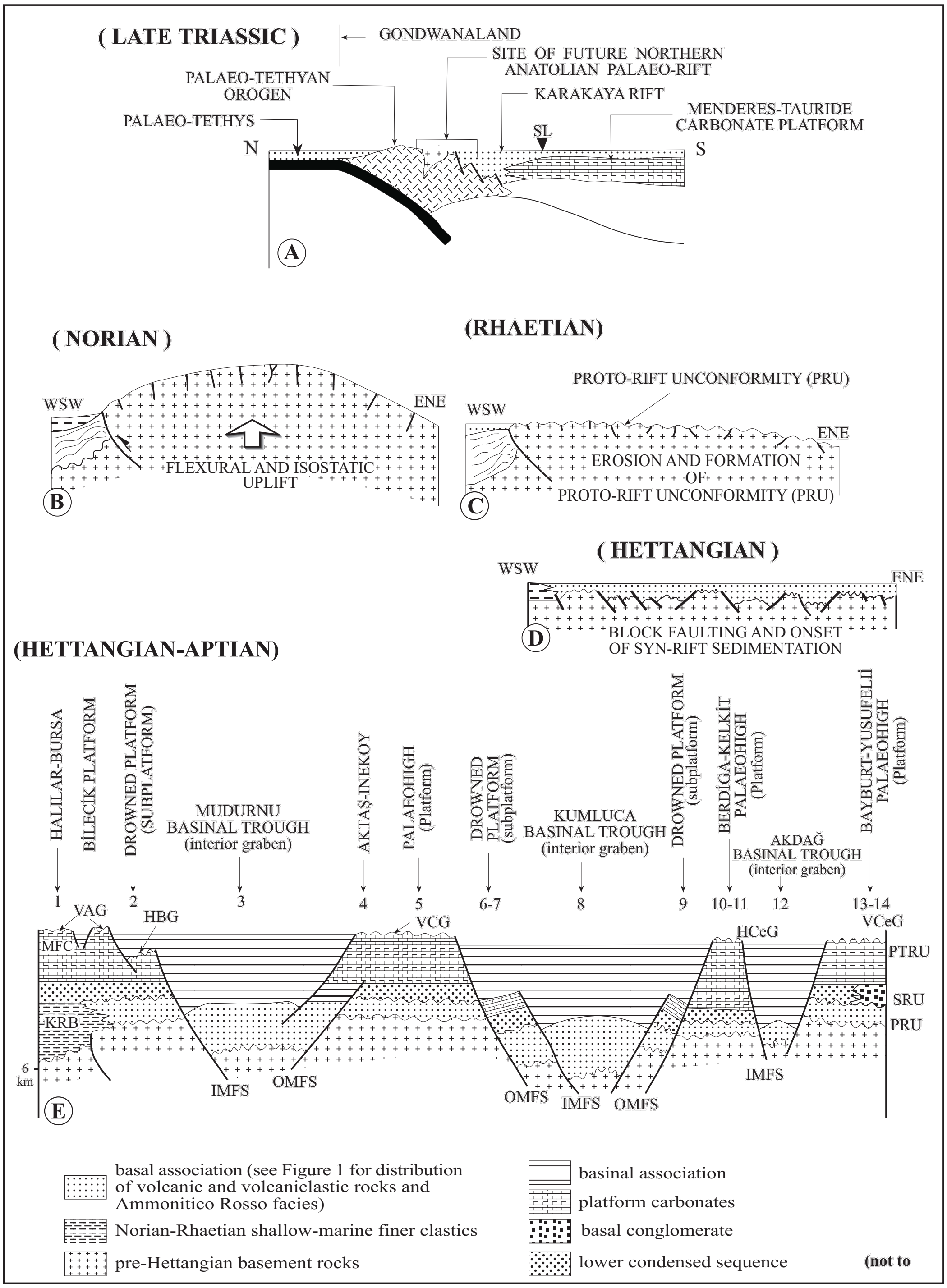

Figure 2. **(A)** Plate tectonic framework showing the future site of North Anatolian palaeorift. SL: sea level; **(B)** uplift and subvertical fracturing; **(C)** formation of proto-rift unconformity; **(D)** block faulting and initial sedimentation of the North Anatolian Palaeorift infill, and **(E)** diagrammatic three-stage graben model depicting the fully developed North Anatolian Palaeorift. HBG- Hauterivian–Barremian gap, HCeG- Hauterivian–Cenomanian gap, IMFS- inner master fault system, KRB- Karakaya remnant basin, MFC- marginal fault complex, OMFS- outer master fault system, PRU- pre-rift unconformity, PTRU- post-rift unconformity, SRU- syn-rift unconformity, VAG- Valanginian-Aptian gap, VCG- Valanginian-Coniacian gap, and VCeG- Valanginian-Cenomanian gap.

of varying size, and consist mainly of serpentinized, sheared and brecciated peridotite, harzburgite, hornblende gabbro, diorite and diabase with minor amounts of basalts (Koçyiğit 1991a).

(e) The fifth basement rock is a tripartite facies association characterizing the Triassic Karakaya rift. It is well exposed at locations 6 and 7 in Figure 1 and consists, from bottom to top, of arkosic clastics, alternations of shallow-marine carbonate, pillow basalts and their pyroclastic equivalents, and a wildflysch. Based on fossil content, this association is considered to be Triassic in age (Koçyiğit *et al.* 1991b; Altıner & Koçyiğit 1993).

Finally, in the Halılar area only (location 1 in Figure 1), the North Anatolian palaeorift fill has a transitional contact relationship with underlying Triassic fine-grained clastics of the remnant Triassic Karakaya rift (Koçyiğit 1990).

North Anatolian Palaeorift Infill

On the basis of depositional settings and facies, the NAPR infill is divided into three basic facies associations: basal association, platform association, and basinal association (3, 4, 6 and 7 in Figure 2E). The basal facies association constitutes the syn-rift infill, whereas both the platform and basinal associations represent the post-rift infill.

The Basal Association– Except for the Halılar area (1 in Figure 1), the basal association rests unconformably on an erosional surface of the pre-Hettangian basement rocks. In contrast, at the top, it passes into the basinal association at locations 3, 8 and 12, while it is overlain unconformably by platform carbonates at the remaining locations (Figure 1).

The basal association ranges from 100 m to 3.8 km in total thickness, and consists mainly of five basic units of varying thickness that have both lateral and vertical transitional contact relationships. These are: (1) basal conglomerate, (2) continental to shallow-marine clastics with intercalation of volcaniclastic rocks, (3) Ammonitico Rosso deposits, (4) volcaniclastic and volcanic rocks, and (5) coal- and gypsum-bearing clastic rocks.

Basal Conglomerate– At the type (2, 5, 6, 7, 8, 11, 12, 13, 14) localities, the NAPR fill starts with a basal conglomerate on the erosional surface of the afore-mentioned basement rocks with angular unconformity or nonconformity (Figure 1). It is an unsorted to poorly sorted polygenetic fan conglomerate and consists of well rounded to sub-angular pebbles and cobbles ranging from a few cm to 1.5 m in diameter. The clasts comprise granite, quartz, marble, Permian to Triassic recrystallized limestone, schist, phyllite, spilitic basalt, serpentinite, metaultramafic rock, chert and radiolarite. These clasts are set in a litharenitic matrix bounded by a ferruginous and siliceous to limy cement.

The basal conglomerate is mostly massive or very thick-bedded. It occasionally displays some sedimentary structures such as cross-bedding, pebble imbrication, graded bedding and mesoscopic growth faults. This conglomerate also contains sandstone lenses and plant fragments and ranges in thickness from a few meters up to 200 m. Although no diagnostic marine or continental fossils have been found within the basal conglomerate, it is followed by shallow-marine clastics or a Ammonitico Rosso facies of late Hettangian–Pliensbachian age (Koçyiğit 1987; Altıner *et al.* 1991; Nicosia *et al.* 1991). It is interpreted as a continental-fan conglomerate deposited by debris flow and braided-river systems.

Continental to Shallow-marine Clastics– These rocks are well exposed in the Bilecik area (Altıner *et al.* 1991) and at type localities 1, 5, 6 and 12 (Figure 1). They start with a coarse-grained arkosic sandstone on an erosional surface of the basement rocks, or conformably follow the basal conglomerate. They comprise alternations of medium- to thick-bedded shallow-marine sandstone, thin-bedded to laminated, green-yellow siltstone, shale, marl, turbiditic sandstone and red mudstone to marl with volcanogenic sandstone intercalations; the latter unit is possibly the first product of rift volcanism in the NAPR. These alternations contain lenses of both calcareous and marly Ammonitico Rosso facies, with rich associations of ammonites (Alkaya 1979; Cope 1991), brachiopods (Ager 1991), bivalves and gastropods (Conti & Monari 1991) belemnites (Doyle & Mariotti 1991), crinoids (Nicosia 1991) and foraminifera (Altıner 1991; Altıner *et al.* 1991); this fauna constrains the age of these shallow marine rocks as Hettangian–Pliensbachian (Koçyiğit 1987; Altıner *et al.* 1991; Nicosia *et al.* 1991). The continental to shallow-marine clastics range in thickness from 100 m to 700 m.

Ammonitico Rosso Deposits– These deposits occur as lenses, a few metres to 250 m long, 30 cm to 70 m in thickness, or as 1-cm to 10-cm-thick beds alternating with shallow-marine fine-grained clastic rocks. Ammonitico Rosso deposits are of three major types: (a) nodular calcareous Ammonitico Rosso, (b) clayey-nodular marly Ammonitico Rosso, and (c) red silty Ammonitico Rosso. The nodular calcareous Ammonitico Rosso also occurs in three sub-facies which are the stromatolitic nodular limestone, bioturbated and bioeroded limestone and true nodular limestone. The nodular calcareous Ammonitico Rosso facies is massive to thick-bedded and grey to pinkish in colour. It shows a transitional bottom contact relationship with the underlying coarse-grained and shallow-marine sandstones, but a sharp to faulted-contact relationship with the overlying relatively deep-marine and fine-grained turbiditic clastic rocks. It is composed of oobiosparite to oobiomicrite and includes karstic features and cracks with iron-rich infill; it lacks evidence of terrigenous influx. All of these characteristics indicate deposition on a shallow plateau or seamount open to the influence of wave action (Bernouilli & Jenkyns 1974; Nicosia *et al.* 1991).

On the other hand, the clayey-marly and silty Ammonitico Rosso deposits occur as lensoidal intercalations within relatively deeper-marine, fine-grained clastics to turbidites which are rich in terrigenous influx. The Ammonitico Rosso has transitional contact relationships with fine-grained marine clastic rocks, and lateral discontinuity with high terrigenous content is attributable to the influence of currents in relatively deeper marine settings (Bernouilli & Jenkyns 1974; Nicosia *et al.* 1991).

As is widely observed throughout the Alpine belt, Ammonitico Rosso deposits are a key horizon within the Liassic basal units across northern Turkey. Based on various fossil assemblages, a Hettangian–Pliensbachian age is assigned to these deposits (Koçyiğit 1987; Altıner *et al.* 1991).

Volcaniclastic and Volcanic Rocks– At the type localities 1, 6 and 12 (Figure 1), volcaniclastic and volcanic rocks are not observed. In contrast, except for the Mudurnu area (3 in Figure 1), these rocks increase in amount and dominate the basal association of the NAPR in an eastward direction (Figure 1). These rocks succeed the sedimentation of the continental basal conglomerate or coarse-grained arkosic sandstone. They consist of alternations of volcanogenic sandstone, tuffite, tuff, volcanic breccia, spilitic pillow basalt, andesitic lava, shale, mudstone, turbiditic sandstone and pelagic carbonate. They also contain coal seams, Ammonitico Rosso lenses, petrified wood and plant-debris intercalations, and are intruded by diabase sills. The total thickness ranges from 50 to 3000 m (Wedding 1963; Nebert 1964; Alp 1972; Seymen 1975; Pelin 1977; Öztürk 1979; Ketin 1983; Gürsoy 1989; Saner 1990; Koçyiğit *et al.* 1991a, b).

The volcaniclastic and volcanic rocks have a transitional contact with underlying continental basal conglomerates and arkosic sandstones. At the top, they either pass upward into the basinal pelagic association, as in the Mudurnu and Kumluca areas (3, 4 in Figure 1), or they are overlain disconformably by platformal successions ranging from coal-bearing clastics to shallow-marine carbonates (2, 5, 7, 9, 10, 11, 13, 14 in Figure 1). Based on both their stratigraphical position and the Ammonitico Rosso deposits included in the volcaniclastic and volcanic rocks, their age ranges from Pliensbachian (or pre-Pliensbachian in the Liassic) to Callovian. For example, in the Mudurnu and Aktaş areas (3, 4 in Figure 1), volcaniclastic and volcanic rocks are overlain conformably by Callovian pelagic carbonates. In contrast, in the Günyarık area (2 in Figure 1), the same volcaniclastics are underlain conformably by the Hettangian–Pliensbachian Ammonitico Rosso deposits, but overlain disconformably by the Callovian shallow-marine platform carbonates. Similar kinds of stratigraphical contact relationships are also observed in the Ladik, Kumluca and Reşadiye areas (7, 8, 9 in Figure 1). All of these observations indicate that the early volcanism began during the Pliensbachian or pre-Pliensbachian stages of the Liassic and continued up into Callovian within basinal areas, such as the Mudurnu and Kumluca troughs (3, 8 in Figure 1).

Petrographical and petrochemical studies carried out on these volcanic rocks by Bektaş *et al.* (1987) indicate a remarkable compositional polarity towards the south from tholeiitic to calc-alkalic through to high-K calc-alkalic to high-K calc-alkalic-alkalic. The polarity and petrochemical characteristics of the volcanic rocks indicate a magmatic origin related to a southward subducting Palaeo-Tethys oceanic crust (Şengör *et al.* 1980; Bektaş *et al.* 1987). This tholeiitic to calc-alkalic magmatism is one of the distinctive characteristics of the rifting phase of a supradetachment basin situated on young orogens or in a back-arc setting (Friedmann & Burbank 1995). Consequently, the petrochemical characteristics of Pliensbachian–Callovian volcanic rocks in Turkey support the supradetachment basin interpretation of the NAPR.

Coal- and Gypsum-bearing Clastics– A continental to shallow-marine sequence occurs at locations 10, 11, 13 and 14 in the eastern part of northern Turkey (Figure 1). This sequence has a short-term erosional contact relationship with underlying volcaniclastic to volcanic rocks at locations 10, 11 and 13, and displays the same contact relationship with both underlying and overlying units at location 14 (Baydar *et al.* 1967).

At locations 10, 11, 13, 14 (Figure 1), volcaniclastic and volcanic rocks are replaced upward by a sequence of gypsum, green shale, marl and iron-rich, limy and cross-bedded sandstones that also includes 1-m-thick coal seams. The total thickness of this sequence ranges from 40 m to 500 m. It passes upward into thick-bedded, shallow-marine and oolitic platform carbonates with the exception of location 14, where it is overlain unconformably by a thick conglomerate (Figure 1).

Palynologic studies carried out on coal samples taken from the Kelkit and Bayburt areas (11, 13 in Figure 1) show that this sequence is Bajocian–Bathonian in age (Ağralı *et al.* 1965, 1966).

The Platform Association

In general, the platform association is characterized by shallow-marine carbonates. It either rests unconformably on an erosional surface of the underlying basal association (2, 5, 6, 7, 9 and 14 in Figure 1) or shows a transitional contact relationship with it (1, 4, 10, 11 and 13 in Figure 1). In the same way, its upper contact is either unconformable or transitional with the pelagic basinal association (Figure 1). Apart from these localities, the outcrops of the platform carbonates also occur in the Sivrihisar-Polatlı-Haymana region underlying the fore-arc sequences of Cretaceous to Eocene age (Gautier 1984; Koçyiğit & Altıner in preparation). The age range of the platform association is from Bajocian in the Halılar area (1 in Figure 1) to Barremian in the Ladik area (7 in Figure 1), and its total thickness varies from 100 m to 1500 m.

Based on lithofacies and biofacies, the platform association is divided into two sequences: a lower condensed sequence, and an upper platform carbonate sequence (see legend in Figure 1B).

Lower Condensed Sequence– The lowermost facies of the condensed sequence changes with distance away from the basinal trough. Near this trough it is a thin volcanic rock, or a maximum 1-m-thick basal conglomerate, but away from the trough it comprises a 1-2-m-thick ammonite-rich stromatolitic and pelletic condensed packstone. The lower condensed sequence starts with one of these basal units and continues upward as 15-cm to 1-m-thick beds of strongly bioturbated, pinkish and iron-rich nodular limestones. The nodular facies is usually composed of pelagic pellet- and oolite-rich packstones and rarely of grainstones. Pellets and oolites are bound by stromatolitic algae resulting in the formation of bindstones or even oncolitic packstones. The main part of the condensed sequence is characterized by cherty dolomite and medium to thick-bedded, dark grey, siliceous limestones composed of pelagic oolite and pelletic grainstones and packstones with abundant thin-shelled bivalves, and crinoid-rich pelletic and intraclastic packstones. In the uppermost part, the condensed sequence becomes less pelagic with the appearance of a *Tubiphytes* facies, and is replaced by the platform carbonates. The condensed sequence ranges from 3 to 350 m in thickness and, based on a very rich macro- and microfaunal-association, its age ranges from Bajocian to early Kimmeridgian (Altıner 1991; Altıner *et al.* 1991).

Stromatolites, pelagic-fossil content (ammonites, *Globochaete, Globuligerina*), and condensed nodular limestones with pelagic oolites and pellets indicate an open marine but shallow-water depositional setting, such as a seamount or pelagic continental plateau, during sedimentation of the lower condensed sequence (Bernouilli & Jenkyns 1974; Jenkyns 1986) where oolites formed under strong wave action.

Upper Platform Carbonates– These carbonate rocks are conformable with the underlying condensed sequence but display both transitional and erosional upper contact relationships with the deep-marine pelagic levels of the basinal association. They range from 100 m to 1100 m in thickness. Detailed palaeontologic studies on these platform carbonates indicate an age range from Kimmeridgian to Aptian (Altıner *et al.* 1991).

The upper platform carbonates occur in two major subsequences: (a) an upward-fining (transgressive) subsequence, and (b) an upward-coarsening (regressive) subsequence. The upward-fining subsequence starts with medium- to thick-bedded, beige to grey limestones. These limestones consist of coral boundstones, bioclastic grainstones, oolitic, flat pebble- and pellet-rich packstones and grainstones, dasyclad algal and pelletoidal wackestones, packstones and ostracod and pellet-rich packstones. Toward the top they pass, in turn, into thin- to medium-bedded cherty limestones and cherts and then to radiolaria-bearing, thin-bedded to laminated argillaceous pelagic carbonates. These upper levels characterize the basinal association. In contrast, the regressive subsequence starts with limestones similar to the fining-upward sequence, but is followed by two distinct horizons. The first horizon, forming the middle part of the regressive subsequence, is characterized by oncolite- and nerineid gastropod-rich grey-pinkish limestones; these are rich in oncolitic, oolitic and pelletic packstone, oncolitic grainstone to mudstone and bioturbated and pelletic mudstones. The second horizon, which constitutes the upper part of the regressive subsequence, is characterized by beige to grey micritic and pelletic limestones containing requienid rudist accumulations at several levels. These latter limestones are composed of various facies including bioturbated mudstone, laminated, pelletic and birdseye mudstone and oncolitic grainstone, mudstone, requienid rudist- and oncolite-rich wackestone. The uppermost part of the regressive subsequence is dominated by meter-scale cycles consisting of stromatolitic and requienid rudist levels.

The regressive subsequence evolves vertically into back-reef deposits characterized by oncolitic and bioturbated nerineid grainstones and packstones rich in foraminifera and dasyclad algae deposited in an intertidal to subtidal environment. The requienid rudist limestones alternate with birdseye-rich pelletic levels forming meter-scale cycles laid down in subtidal, intertidal and supratidal zones. Similar regressive successions have been previously described by Colacicchi *et al.* (1975) from Italy. The cycles forming the uppermost part of the regressive subsequence resemble the “grainy beach cycle” deposits described from the Jurassic of the Paris basin by Purser (1975). The regressive subsequence of the upper platform carbonates also displays a well-developed, short- to long-term palaeohigh character. Starting from the Valanginian, the regressive subsequence begins to emerge and persists as a palaeohigh until succeeded by sedimentation of deep-marine facies of the basinal association on an erosional surface. Accordingly, near the uppermost contact of the regressive sequence, the blue-green algal mudstones are cut by an erosional surface and overlain disconformably by a pelagic facies of the basinal association with ages varying from Aptian to Santonian. This erosional contact is marked by the occurrence of karstic features, neptunian dykes consisting of a reddish sediment rich in iron, silica and calcite indicative of short- to long-term breaks in sedimentation. A lithostratigraphical and biostratigraphical case study of the well-developed palaeohigh of Aktaş-İneköy (Figure 1) is described here.

Aktaş-İneköy Palaeohigh– Although the Aktaş and İneköy sections (4 and 5 in Figure 1) represent different palaeogeographic domains within the Jurassic–Early Cretaceous evolution of northern Anatolia, they are conceptually gathered together in this study into an east–west-trending, two-dimensional graben model (Figure 2). The Aktaş succession was previously defined as one of the important sections representing the Aktaş-Sekinindoruk high bordered from north, west and east by the Mudurnu trough (Koçyiğit 1987; Koçyiğit *et al.* 1991a). The İneköy section is possibly located on a more southerly margin of this basinal trough (Figure 1), representing a similar kind of palaeohigh. The İneköy outcrop occurs as an inlier surrounded by Tertiary sedimentary cover-rock assemblages divided into several formations, approximately 35 km northwest of Ankara (Figure 3).

The succession in the İneköy section (5 in Figure 1; Figure 4), overlies basement rocks of pre-Jurassic age (metamorphic rocks of the Pazaryeri structural complex) and is basically characterized by two main sedimentary rock assemblages, namely a basal association and a platformal association, both displaying the characteristics of the NAPR infill.

Two formations have been distinguished in the basal association (Figures 3 and 4). The lower-lying Yılgınçay Formation measures about 18 m in thickness and consists

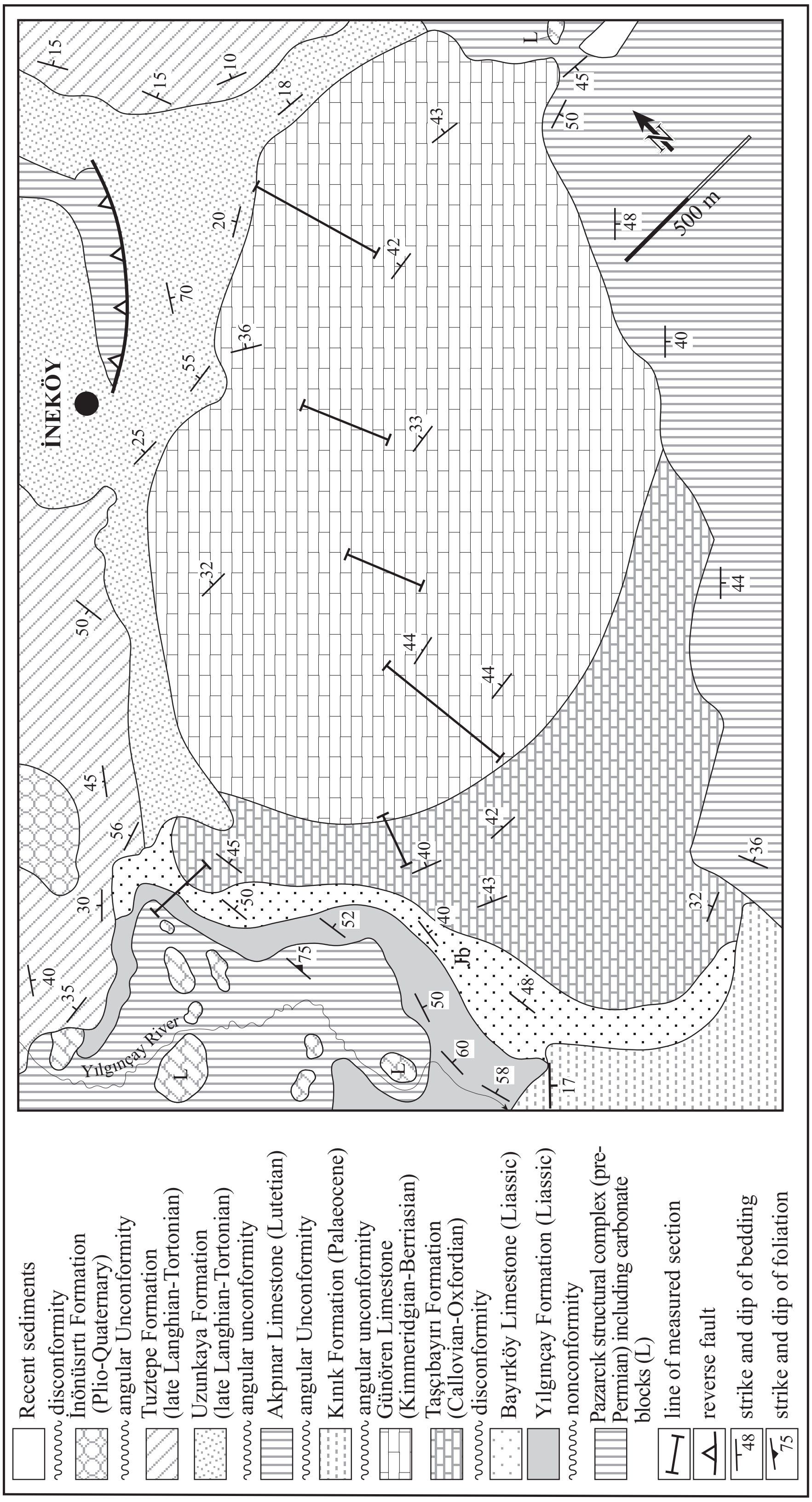

Figure 3. Geological map of the İneköy Jurassic–Lower Cretaceous inlier.

of thin-bedded, yellow to brown siltstones and sandstones. These clastics are of volcanogenic origin and contain abundant volcanic glass and tuffaceous material. They are newly discovered in Ankara region and are interpreted as the first product of volcanic activity in the NAPR. The upper part of the formation is undoubtedly marine in origin since samples collected at level 1 (Figure 4) yielded sponge spicules, radiolaria and a poor foraminiferal assemblage indicating a post-Triassic age. The upper part of the basal association is represented by the Bayırköy Formation, a well-known lithostratigraphic unit from northwestern Anatolia (Altıner *et al.* 1991; Koçyiğit *et al.* 1991a, b). This formation begins with yellow-brown, crinoid-rich sandstones containing volcanic quartz, feldspar and metamorphic-rock fragments set in a matrix of volcanic origin (level 2, Figure 4). These levels grade into finer-grained, thinly laminated sandstones and are intercalated with sandy oolitic packstones, including many macrofossil fragments, and an important foraminiferal assemblage including *Involutina liassica* (level 3, Figure 4). The remaining part of the Bayırköy Formation (levels 4-6, Figure 4) consists of sandstones with yellowish-brownish pebbly layers containing clasts up to 4 cm in diameter. Marine fossils, oolites and bioclastic material are present in these levels, but usually subordinate to the siliciclastic material comprising quartz, feldspar, mica, metamorphic-rock and granitic rock fragments. The Liassic age, confirmed by the presence of *Involutina liassica* for the Bayırköy Formation, is also significant for dating the beginning of volcanic activity in northern Anatolia. *Involutina liassica*, indicating a maximum Pliensbachian age calibrated by ammonites (Altıner *et al.* 1991), occurs in level 3 (Figure 4) overlying volcanogenic sandstones of the Yılgınçay Formation. This suggests that volcanism commenced during the Pliensbachian or pre-Pliensbachian stage of the Liassic and modifies all previous age assignments for the earliest age of the volcanic activity in northwestern Anatolia (Saner 1980; Koçyiğit *et al.* 1991a, b; Altıner *et al.* 1991), which suggested that it began during the Dogger.

The platform association in the İneköy section is composed entirely of carbonate rocks which rest unconformably on the basal association of Liassic age. The unconformity surface is a 1-m-thick, yellowish oxidized zone containing thin travertine deposits. Two distinct formations have been recognized in the platform association (Figures 3 and 4). The Taşcıbayırı Formation (Altıner *et al.* 1991), which represents mostly the lower condensed sequence of the platform association, measures 73 m and begins at the base with thick-bedded, white to yellow limestones of algal and sponge boundstone and pelagic oolite- and pellet-bearing packstone to grainstone facies (levels 7-8). The faunal content is identical to that described from the type locality of the Taşcıbayırı Formation (Altıner 1991; Altıner *et al.* 1991) in the Bilecik region, and a Callovian–Oxfordian age has been assigned to this lower part of the formation. The rest of the Taşcıbayırı Formation (levels 9-12) is generally composed of white, thick-bedded to massive limestones containing chert horizons. These levels are oncolitic, intraclastic, pelletoidal and bioclastic packstones to grainstones; they contain *Protopeneroplis striata, Mesoendothyra izjumiana* and *Tubiphytes morronensis*, indicating a record of the Kimmeridgian stage (Altıner 1991) above the condensed Callovian–Oxfordian limestones of the Taşcıbayırı Formation.

The succeeding Günören Limestone which represents the upper platform carbonates of the platform association, measures 910 m in thickness and is broadly cyclic in nature. In the lower half of the formation, levels 12 to 25 correspond to the regressive part of a cycle, developed over a transgressive phase of the platform represented in the lowermost part of the Taşcıbayırı Formation and record a sudden drowning. These levels are composed of medium- to thick-bedded, yellow-pink and white, locally dolomitic, oncolitic, pelletoidal or intraclastic mudstones or wackestones and algal boundstones. Fossils are not diverse, but the presence *of Mesoendothyra izjumiana, Otaina* ? sp., *Paravalvulina* sp. indicates that the succession still belongs to the Kimmeridgian. The overlying next broad cycle in the Günören Formation occurs between levels 26 and 36. The lower part of this cycle (levels 26 to 30, Figure 4) consists of medium- to thick-bedded, high-energy limestones of oncolitic, intraclastic, pelletoidal packstone to grainstone facies, rarely intercalated with wackestones. Upward, the succession grades into relatively low-energy limestones (levels 31-36), comprising basically the partially brecciated, oncolitic and dasyclad algal (*Clypeina*-rich) wackestones. These levels are locally extremely karstified and include vadose silt infillings which indicate subaerial exposure. The fossil data (Figure 4) indicate that the succession in this cycle belongs to the Kimmeridgian and possibly to the lower part of the Tithonian stages. The karstified surfaces and

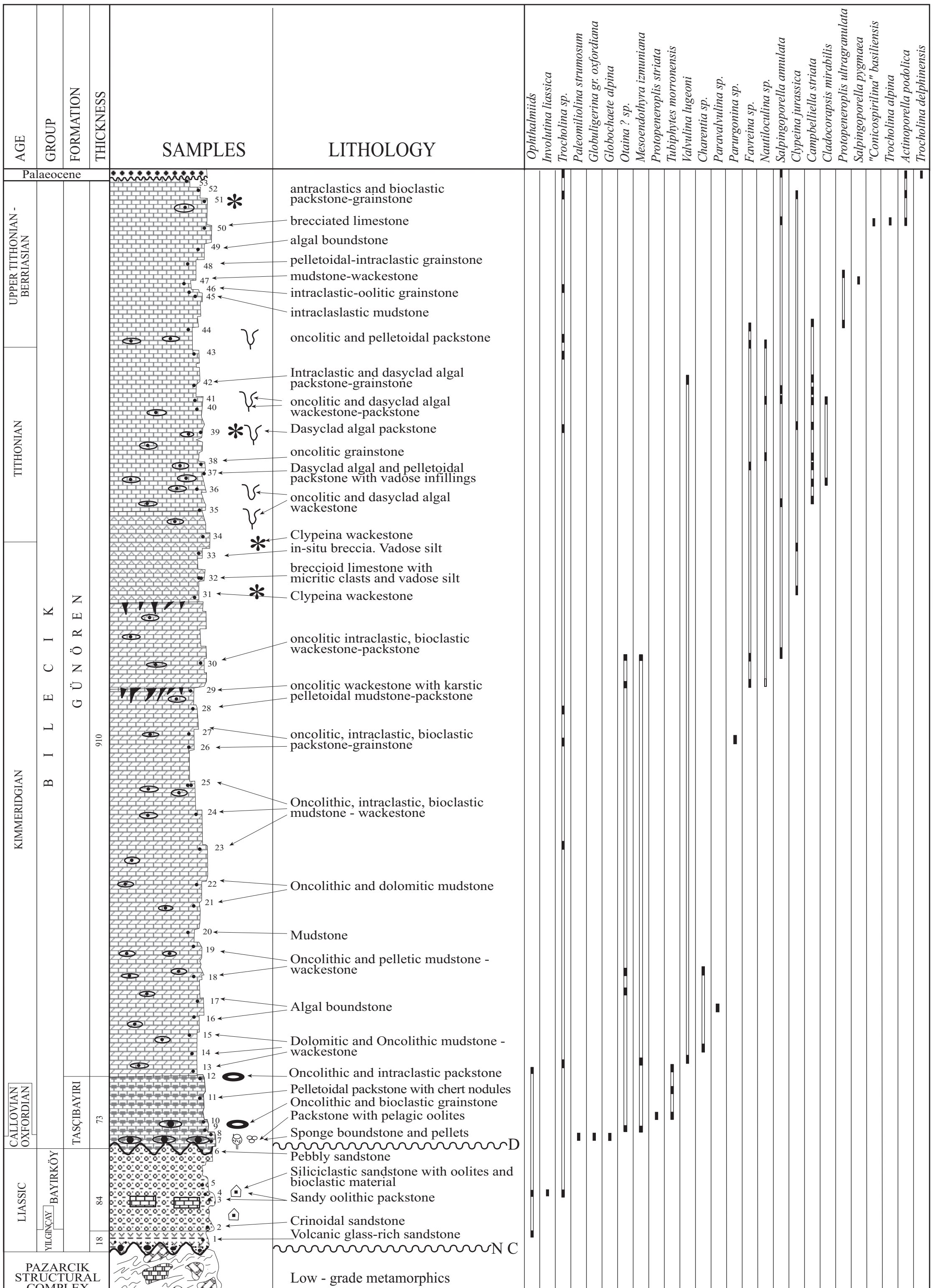

Figure 4. İneköy combined measured stratigraphic section depicting both the lithofacies and biofacies of the Aktaş-İneköy platform carbonates (see Figures 1 & 3 for its location). D- disconformity, and NC- nonconformity.

brecciated levels, which correspond broadly to the Kimmeridgian–Tithonian boundary, are possibly related to repeated eustatic sea-level drops in the early Tithonian (138 My and 136 My, see Haq *et al.* 1987) as in the case of Tauride carbonate platform representing the southern margin of the northern Neo-Tethys ocean (Altıner *et al.* 1999).

Following these karstified, subaerially exposed levels, the rest of the Günören Limestone (levels 37-53, Figure 4) is basically characterized by high-energy limestones indicating the beginning of a new cycle. These white to yellow-white and medium- to thick-bedded levels are composed of oncolitic, dasyclad algal, intraclastic and pelletoidal packstones and grainstones (levels 37-44, 46, 48, 51-53), with rare intercalations of intraclastic mudstones to wackestones (levels 45, 47) and algal boundstones (level 49). Above the Kimmeridgian stage, these levels contain *Clypeina jurassica, Campbelliella striata* and *Protopeneroplis ultragranulata* and are of Tithonian to Berriasian age (Figure 4). The presence of *Clypeina jurassica* in the uppermost part of the section restricts the age of the unit at the top to the early Berriasian. The Günören Limestone above the lower Berriasian strata is truncated by an unconformity surface and is overlain by Palaeocene clastics.

The Basinal Association

This is an approximately 1.6-km-thick sedimentary prism dominated by redeposited sediments. This association conformably overlies either volcaniclastic or volcanic and shallow-marine clastic rocks of the basal association within troughs such as those of Mudurnu, Kumluca and Akdağ (3, 8, and 12 in Figure 2E), or platform carbonates along their fault-bounded margins (1, 6, 7, and 9 in Figure 2E), and ranges from Callovian to Aptian in age.

The basinal association is characterized by a thinning- and fining-upward sequence. From bottom to top, it consists of four lithofacies: (1) breccias, (2) thick- and thin-bedded turbidites, (3) basin plain sediments, and (4) bioturbated marl-porcelaneous micritic limestone alternation (Altıner *et al.* 1991).

Breccias- Two types of breccia are recognized, namely unsorted breccias and megabreccias. The unsorted breccias are restricted to the base of the basinal association, or very close to it, and are characterized by poorly sorted diverse lithologies. Components are mostly angular but all degrees from subrounded to rounded are present. The size ranges from sand to megablocks, up to tens of metres in diameter. A number of unsorted breccias are exposed near the base of the basinal association. Two of these are well exposed. The first occurs very close to the margin and is grey-black or yellow-pinkish, with massive to weakly-formed thick-bedding (80 cm–2 m). It consists of subrounded to angular clasts of grainstone, micritic, pelletic, intraclastic, oncolitic limestone and bioclastite containing gastropods, bivalves, algae, corals and foraminifera. The lithofacies and biofacies evidence indicates that these deposits were shed into troughs from the nearby fault-bounded submarine highs (shallow-marine pelagic plateau). This breccia also includes volcanic glass and spilitic basalt fragments, and the whole of this clastic matrix is tightly bound by a dolomitic and siliceous limy cement. The breccia is 180 m thick and conformably underlain by the 3-m-thick spilitic basalt. The uppermost level is very rich in gastropod and bivalve remains and is conformably overlain by a 3-m-thick red conglomeratic breccia and a 20-m-thick spilitic basalt. Radiometric dating studies on the bivalves by the $^{87}Sr/^{86}Sr$ method have yielded an age of 151 Ma indicating a Kimmeridgian age (Altıner *et al.* 1991). The foraminiferal and algal association supports the assignment of a Kimmeridgian age.

A second example of the unsorted breccias occurs within the trough farther from the margin where it is found as an intercalation within a condensed and deep-marine sequence. This sequence is composed of, from bottom to top, alternations of 2-m-thick spilitic basalt, thin- to medium-bedded wackestone rich in radiolaria and chert nodules, jasper, green tuff, volcanogenic sandstone with abundant current-oriented belemnites, pelletic mudstone rich in thin-shelled bivalves, thin and unsorted breccia, 3-m-thick spilitic basalt, 8-m-thick wackestone and the unsorted breccia horizon which is 18 m thick. The breccia consists of subrounded to angular clasts of Kimmeridgian grainstone with *Tubiphytes*, packstone containing pelagic oolites, mudstone and spilitic basalt set in a tuffaceous matrix. The clasts range in diameter from 1 cm to 20 cm and lithofacies and biofacies of the clasts indicate that they were shed into the trough from a fault-bounded submarine high of platform carbonates. This unsorted breccia is conformably followed upward by a 20-m-thick spilitic basalt and thin- to thick-bedded pelagic limestones interbedded with turbiditic laminae. Foraminiferal and macrofossil associations of the condensed sequence, including unsorted breccias, have yielded a Callovian–Kimmeridgian age (Altıner *et al.* 1991).

In contrast, chaotic megabreccias, which range in thickness from 0.5 m to 100 m, occur concordantly within basinal sediments of Tithonian–Barremian age, and are intercalated with both coarser- and finer-grained calciturbidites as well as bioturbated marls and limestones. Megabreccias are poorly organized and do not cut into the underlying sequences. They consist of very coarse, poorly sorted, angular to moderately rounded components up to 10 m in diameter. The matrix is composed of marl and finer-grained clasts of similar composition. The general appearance of the megabreccias is chaotic and structureless.

One of the well-developed examples of megabreccia occurs near the eastern margin of the Mudurnu trough (3 in Figure 2E) and comprises a monogenetic limestone breccia within a 100-m-thick zone of regularly bedded carbonate turbidites. Components include various carbonate facies such as packstone, grainstone, wackestone and reefal bindstone with *Tubiphytes, Lithocodium*, gastropods and some terrigenous material. Clasts range in size from a few cm to blocks 10 m in diameter. The matrix is composed of marl and finer-grained clasts of the same composition. Both the foraminiferal and algal associations have yielded a late Berriasian–early Valanginian age (Altıner *et al.* 1991).

A second example of megabreccia occurs as six megabreccia horizons within an 80-m-thick zone of finer-grained carbonate turbidites of late middle Tithonian age at the same locality. These horizons range in thickness from 0.5 m to 2 m and consist of subrounded to angular clasts of bindstone with corals, wackestone with sponge spicules, bioclastites with *Tubiphytes* and mudstone set in an argillaceous-arenitic matrix. Components are also monogenetic and range in diameter from 1 mm to 25 cm. This zone of megabreccias is overlain conformably by a finer-grained carbonate turbidite containing *Zoophycus*.

In addition to the afore–mentioned breccias, a number of isolated limestone olistoliths also occur within the finer-grained carbonate turbidite, bioturbated marl and tuffaceous siliciclastic rocks. They range in size from a few cubic meters to tens of cubic meters and consist of various carbonate facies; these are mainly bioclastic grainstones, packstones with pelagic oolites, intraclastic packstone and bindstone with *Lithocodium*, sponges and brachiopods. The matrix is a slumped and chaotic mixture of siltstone, blue and bioturbated marl, shale, tuffaceous wackstone and micritic finer-grained carbonate turbidite in which olistoliths display more than one bed-parallel alignments and can be traced laterally for at least 2 km. The age of the olistoliths ranges from Callovian to Valanginian, and their litho- and biofacies indicate that they have been derived from fault-bounded submarine highs shedding into adjacent deep-marine, basinal-association troughs (Koçyiğit *et al.* 1991a).

Thick- and Thin-bedded Carbonate Turbidites– The most common rocks filling fault-bounded troughs such as those of Mudurnu, Kumluca and Akdağ (3, 8, and 12 in Figure 2E) are carbonate turbidites. They occur in the Oxfordian–Barremian interval and consist of thick- and thin-bedded carbonate turbidites marked by occurrences of graded bedding, mostly in the $T_{a,b}$ and T_e facies of the Bouma sequence. These are also frequently accompanied by slump and sole structures. Carbonate turbidites contain shallow-marine carbonate clasts up to 5 cm in diameter, and terrigenous materials such as quartz and feldspar grains, volcanic glass, basaltic clasts, and shell fragments of various fossils, such as *Tubiphytes, Lithocodium*, gastropods, bivalves, ammonites, echinoids, sponges and brachiopods. All of these lithoclasts and bioclasts were derived from the nearby unstable shelf edge or a fault-bounded, shallow-marine pelagic plateau, and transported down the slope or fault-induced escarpment into the finer basinal facies by high-energy turbidity currents.

Thick- and thin-bedded carbonate turbidites occur in well-developed thinning and fining-upward sequences up to 1100 m in thickness. One such sequence is well preserved around the Aktaş-İneköy carbonate platform and conformably overlies a 7-m-thick megabreccia horizon containing metre-sized olistoliths of shallow-marine origin. It continues upwards with a 90 m thick, radiolaria-, belemnite- and ammonite-rich, thin- and medium-bedded limestone in wackestone and mudstone facies. The sequence continues farther up in a thick turbiditic cycle with a lower part composed of thin- and thick-bedded limestones displaying graded bedding and sole sedimentary structures, and an upper part having finer-grain size with limestones containing chert nodules and green tuffaceous material. In this cycle, carbonate turbidites are usually composed of silicified packstones, which contain up to 30% noncalcareous material consisting of feldspar, quartz and volcanic-rock fragments. Calcareous elements are transported oolites, micritic or reefal clasts and larger foraminifera. The uppermost part of the sequence is again composed of carbonate turbidites alternating with six distinct monogenetic limestone breccias up to 2 m in thickness. The uppermost beds of the sequence contain *Zoophycus* and are gradually overlain by micritic limestones of the basin plain. The macrobiota and microbiota studied by Altıner *et al.* (1991) yield an Oxfordian to late Tithonian age span for this sequence.

Lithofacies of this sequence can be compared to the redeposited carbonate facies models of Stow (1986) and correspond to the "calcidebrite", "calcirudite-calcarenite" and "calcarenite-calcilutite" types. These redeposited carbonate facies occur either away from the platform margins or on the slopes of seamounts (e.g. Mullins & Neumann 1979). The horizon with *Zoophycus*, recognized in the uppermost beds of the sequence, is the trace-fossil-rich turbidite deposit developed on the basin slope (Seilacher 1967).

Basin Plain Sediments– These occur near the axial part of deep-marine troughs and form the distal parts of the thin- and thick-bedded carbonate turbidites. They consist of finer-grained and thin-bedded turbidites which are interbedded with bioturbated marls and limestone and contain pebbly mudstone horizons in places.

Bioturbated Marl and Limestone Alternation– These marls are green, blue and yellow and commonly form massive beds up to 3 m. However, they are also thin–bedded to laminated and the transition from marls to limestones is gradual. Limestone beds are separated by massive marl horizons and contain chert nodules. They include yellow, pinkish or red limestone beds which may be slightly nodular and contain more pelagic fossils. The marl and micritic limestone alternation is characterized by a lack of redeposited sediments and current structures indicating deposition in a low-energy environment.

In northern Turkey, the bioturbated marl and micritic limestone alternation occurs in two positions within the sedimentary column. First, they overlie shallow-water platform carbonates with a sharp contact, which may be a short- or long-term intra-platform gap (mostly paraconformity) marked by karstic features and vertical neptunian dykes rich in iron-bearing sediment. Second, they overlie conformably the carbonate turbidites of the basal association. In addition, in the upper part of the bioturbated marls and micritic limestone alternations, the occurrence of breccias and slumped sediments disappears, whereas radiolarians and pseudoplanktonic bivalves become dominant. This documents a progressive drowning of the submarine highs and indicates that a widespread pelagic environment was becoming established at the end of the rifting phase (Eberli 1987).

In northern Turkey, the uppermost and youngest lithofacies of the basinal association is the bioturbated marl and micritic limestone alternation. It is well preserved in areas within and adjacent to the Mudurnu, Kumluca and Akdağ deep-marine troughs (2, 3, 6, 7, 8, 9 and 12 in Figure 2E). One of the well-preserved sequences of the bioturbated marl and micritic limestone alternations is observed along the western, fault-bounded margin of the Mudurnu trough (2 and 3 in Figure 2E) where the alternations of bioturbated blue to green marl and micritic limestone overlie subtidal to intertidal carbonates of the submarine high following an abrupt facies change (Altıner *et al.* 1991). In this area, the sequence consists of two major lithologic types. The lower one consists of grey-pink limestones with chert nodules. This level begins with crinoid and bryozoan-rich, glauconitic wackestones and packstones and continues upward with argillaceous wackestones and mudstones containing *Globochaete*, radiolaria and planktonic foraminifera. The uppermost lithologic unit comprises an alternation of marl and micritic limestone with chert nodules, containing abundant *Nannoconus*, planktonic foraminifera, radiolaria and belemnites. Based on benthic and planktonic foraminifera and nannofossils, a Valanginian to late Aptian age has been assigned to this 280-m-thick uppermost unit of the basinal association (Altıner 1991; Altıner & Özkan 1991; Altıner *et al.* 1991; Özkan 1993; Özkan-Altıner 1996).

Analogous kinds of basinal facies including breccias, thick-bedded turbidites, thin-bedded turbidites, basin-plain sediments and marl and micritic limestone alternations have been reported from the continental margin of the Jurassic Tethys ocean in the eastern Alps of

Switzerland. This assemblage has been interpreted to have been deposited in a rift-basin between tilted fault blocks (Eberli 1987).

Structural Outline

The main structural elements of the north Anatolian palaeorift (NAPR) are the interior grabens, the subplatforms, the platforms, the rift-related unconformities, and the rift-related faults.

Interior Grabens

The interior grabens are bounded by the inner fault systems on both sides, and occupied by both the lower basal and upper basinal associations. The best-studied examples are the Mudurnu, Kumluca and Akdağ grabens (Figure 2E) (Altıner *et al.* 1991; Koçyiğit 1991a; Koçyiğit *et al.* 1991a).

Subplatforms

The subplatforms are drowned platforms bounded by the outer fault system on the platform-ward side and by the inner master fault system on the interior graben-ward side. They are therefore highly faulted. They are located at one or both margins of the interior grabens and are composed of shallow-water platform carbonates overlain conformably and/or unconformably by the finer-grained basinal association. Good examples are the Günyarık, Hasanoğlan, Ladik and Reşadiye subplatforms (2, 6, 7 and 9 in Figure 2E) (Seymen 1975; Öztürk 1979; Koçyiğit 1987; Altıner *et al.* 1991; Koçyiğit *et al.* 1991a).

Platforms

The platforms are very broad and nearly flat areas made up of shallow-water carbonates and characterized by moderate fault activity and subsidence (Nottvedt *et al.* 1995). Therefore, they may emerge partly or entirely, and persist as palaeohighs over short or long periods of time. The best examples are the Halılar-Bursa-Bilecik, the Aktaş-İneköy, the Berdiga-Kelkit and the Bayburt-Yusufeli platforms (Figure 2E) (Baydar *et al.* 1967; Rojay 1985; Koçyiğit & Altıner 1990; Altıner *et al.* 1991; Koçyiğit *et al.* 1991a).

Rift-related Unconformities

In general, three groups of rift-related unconformities occur during the rift basin evolution as predicted by the three-stage graben model of Nottvedt *et al.* (1995). These are, from oldest to youngest, (a) the proto-rift unconformity, (b) the syn-rift unconformity and (c) the post-rift unconformity. According to Nottvedt *et al.* (1995), the proto-rift unconformity forms by erosion of a rift-related dome, the syn-rift unconformity by erosion of uplifted fault blocks, and the post-rift unconformity by erosion of the entire rift structure. Accordingly, the proto-rift infill is deposited during early flexural subsidence, the syn-rift infill during active stretching, and the post-rift infill during the thermal subsidence phase.

Proto-rift Unconformity– The Palaeo-Tethyan orogen resulted from the southerly subduction of the Palaeo-Tethys Ocean was isostatically and flexurally uplifted during Permo-Triassic time (Figure 2A, B). Accordingly, it was eroded over a broader area throughout northern Turkey until Hettangian time during which a regional erosional surface developed resulting in a stratigraphic gap in the geologic history of northern Turkey; it is referred to here as the proto-rift unconformity (Figure 2C). This proto-rift unconformity is marked by a continental basal conglomerate up to 200 m thick, atop an erosional surface of pre-Hettangian basement rocks. These polygenetic basal conglomerates are the lowermost basal facies of the NAPR infill (Figure 2D) and, as noted earlier, except for the Halılar, Mudurnu and Reşadiye areas, it is well-exposed across northern Turkey (Alp 1972; Yılmaz 1972; Öztürk 1979; Görür *et al.* 1983; Koçyiğit 1987, 1989; Gürsoy 1989; Koçyiğit *et al.* 1991a). The basal conglomerate passes upward into shallow-marine clastics and an Ammonitico Rosso facies deposited by a transgressing sea. This regional late Hettangian transgression initiated the syn-rift sedimentation of the NAPR infill (Figure 2D).

Syn-rift Unconformities– Although the nature, duration and number of unconformities in the eastern part of the NAPR are still poorly known, they have been determined precisely by the detailed stratigraphical, palaeontological and radiometric dating studies of Altıner *et al.* (1991) and Koçyiğit *et al.* (1991a) in its western half.

Within the 6.2-km-thick infill of the NAPR, two groups of relatively long-term, regional unconformities were detected (Figure 4). The older first group of unconformities represent the syn-rift unconformity formed by erosion of uplifted fault blocks based on the three-stage graben model (Nottvedt *et al.* 1995). These are the Toarcian–Bathonian gap (TBG), the Toarcian–Callovian gap (TCG), the Toarcian–Aalenian gap (TAG), the Aalenian gap (AG) and the Callovain gap (CG). The latter three unconformities are still poorly understood. In contrast, the TBG and TCG are better known and separate the underlying Hettangian–Pliensbachian basal association from the overlying Callovian and/or younger shallow-marine platform carbonates (Figures 2E and 4). Along their boundary, a 0.2–0.8-m-thick basal conglomerate and some iron-rich karstic features up to 15 cm in thickness, including travertine and oxidized surfaces, occur. The polygenetic basal conglomerate consists of pebbles derived directly from the underlying basal association. This thin sequence of basal conglomerate and travertine indicates an erosional period spanning from Toarcian to Callovian or Oxfordian. It is overlain conformably by thick-bedded cherty limestone with pelagic oolite (lower condensed sequence) of the basinal association.

Post-rift Unconformity– The second group of stratigraphic gaps have the nature of paraconformities and represent the post-rift unconformity. These are restricted to the intra-platform highs, and their time span ranges from the Valanginian to the end of the Aptian and possibly to the Palaeocene (PH in Figure 4). Among these unconformities are the Valanginian–Aptian gap (VAG), the Hauterivian–Barremian gap (HBG), the Valanginian–Coniacian gap (VCG), the Aptian–Palaeocene gap (APG), the Aptian–Albian gap (AAG), the Hauterivian–Cenomanian gap (HCeG) and the Hauterivian–Turonian gap (HTG). During these gaps, the platforms of shallow-water carbonates were totally emerged and persisted as the palaeohighs inside the NAPR (Altıner *et al.* 1991; Koçyiğit *et al.* 1991a). These long- and short-term gaps or paraconformities and/or disconformities are best evidenced by the occurrence of karstic features such as yellow-reddish travertine, breccia and vertical neptunian dykes; the latter are up to 2 m deep and 1–mm to 30–cm wide, and infilled by iron oxide-rich sediment. These disconformities are also marked by a clear contrast in lithofacies and biofacies on both sides of the gaps, sharply separating underlying massive to thick-bedded and shallow-water platform carbonates from overlying thin-bedded, yellow to red, radiolaria-bearing deep-marine cherty carbonates which are mostly the uppermost facies of the basinal association.

Rift-related Faults

Throughout northern Turkey, another distinctive record of the NAPR is seen in the extensional faults. Based on age, size and position, three groups of rift-related faults are observed: older growth faults, outer and inner fault systems, and a marginal fault complex (Figure 2E). They were previously mapped and defined in detail by Altıner *et al.* (1991) and Koçyiğit *et al.* (1991a).

The first group of faults are high-angle, outcrop-scale normal growth faults restricted to the lowermost facies of the basal association (Koçyiğit *et al.* 1991a). These faults are best exposed within the mid-Hettangian conglomerates-arkosic sandstones, and the upper Hettangian–Pliensbachian shallow-marine clastics to nodular limy Ammonitico Rosso facies. These faults range in size from 1 m to 5 m and cut across the uppermost parts of the basement rocks and the lowermost continental to shallow-marine facies of the overlying basal association. They record the first-phase of faulting and dominate both the proto-rift and also the incipient syn-rift stages predicted by the three-stage graben model of Nottvedt *et al.* (1995).

The second group of faults belong to outer and inner master fault systems and are map-scale, low-angle normal faults (Koçyiğit *et al.* 1991a). They occur within a wide zone (subplatform) and determine the platform-interior graben boundary (IMFS and OMFS in Figure 2E). They cut across both the older basement rocks and the whole of the rift infill. Their existence is best evidenced by the following field observations: (a) a juxtaposition of shallow-marine platform carbonates and finer-grained basinal association along a disturbed and chaotic zone without a transitional facies between them; (b) this chaotic and disturbed zone consists mainly of very thick (up to 180 m) unsorted megabreccias and broken formations; (c) both the platform carbonates and basinal association are of the same age, however the basinal association is much thicker than the platform carbonates; (d) adjacent to the margins of interior grabens in particular, the pelagic basinal association contains abundant syn-depositional features such as megabreccias, giant olistoliths, slump folds, slumped and broken formations and coarse-grained carbonate turbidites which may record fault-induced gravitational instability in the depositonal setting, and (e) although the sedimentation within the interior grabens is continuous a number of short (and a lesser number of long-term) stratigraphic gaps (disconformities and paraconformities) are included in the platforms which are also distinctive records of fault-induced uplift and erosional processes (Koçyiğit *et al.* 1991a). As noted earlier, the same features and facies have also been observed at the continental margin of the Jurassic Tethys ocean in the eastern Alps and have been interpreted as the product of deposition in a rift basin shaped by tilted fault blocks (Eberli 1987).

The second group of faults result from the second- and third-phases of faulting, and they dominate both the syn-rift and post-rift stages of the three-stage graben model (Nottvedt *et al.* 1995).

The third group of faults are also map-scale, low-angle normal faults. These faults are restricted to the platforms, and determine the margins of smaller and younger intra-platform grabens. They are younger than both the outer and inner master fault systems, and are evidenced by more-or-less the same features mentioned above. They can be assigned to a marginal fault complex (MFC in Figure 2E) according to the three-stage graben model of Nottvedt *et al.* (1995).

Discussion and Conclusion

The data presented in this paper support the following geologic history for the tectonostratigraphic evolution of the NAPR.

With the exception of the Halılar area, a new depositional setting began to develop on the erosional ruins of the Palaeo-Tethyan orogen (Şengör 1979; Şengör *et al.* 1980; Yılmaz *et al.* 1997) throughout northern Turkey during Hettangian time (Figure 2A, B, C). This is indicated by a regional proto-rift unconformity and basal continental-fan conglomerates to arkosic clastics which are cut by outcrop-scale, normal growth faults resulting from the first-phase faulting (Figure 2E) (Koçyiğit *et al.* 1991a). These basal continental clastics pass upward into shallow-marine clastics with intercalations of both volcaniclastic rocks and various Ammonitico Rosso facies of late Hettangian–Pliensbachian age (Altıner *et al.* 1991; Nicosia *et al.* 1991). With the exception of the Halılar-Bursa, Hasanoğlan and Akdağ areas (1, 6 and 12 in Figure 1), the Ammonitico Rosso facies-bearing clastics are succeeded predominantly by the products of bimodal (basaltic to andesitic) volcanism (Tokel 1983; Bektaş *et al.* 1987; Yılmaz *et al.* 1997). In the Mudurnu, Aktaş, Reşadiye and Berdiga areas (3, 4, 9 and 10 in Figure 1), the lowermost boundary of the volcanic rocks cannot be observed due to the young sedimentary cover and active faulting. However, newly obtained stratigraphical and palaeontological data from the Günyarık and İneköy areas (2 and 5 in Figure 1) indicate that the volcaniclastic and volcanic rocks in the Mudurnu, Aktaş, Reşadiye and Berdiga areas must be underlain by the Ammonitico Rosso facies-bearing clastics. Thus, the initiation of rift-related volcanism in the NAPR likely occurred during the Pliensbachian or pre-Pliensbachian stages of the Liassic. This assignment for incipient volcanism correlates well with the first-phase of faulting in the NAPR. Outside of the Halılar, Mudurnu, Aktaş, Kumluca and Akdağ districts (3, 4, 8 and 12 in Figure 1), incipient volcanism and the first phase of extensional faulting were followed by regional emergence. This is indicated by the Toarcian–Bathonian gap corresponding to the syn-rift unconformity, which separates the post-Bajocian carbonates from the underlying Hettangian–Pliensbachian basal association. This regional emergence occurred away from the Halılar, Mudurnu, Aktaş, Kumluca and Akdağ areas and is a record of second-phase faulting which divided the basement rocks and their newly deposited younger cover (the basal association of the NAPR) into several blocks, resulting in an irregular tectonomorphology with probable high relief.

Starting most likely from late Pliensbachian time, the lowlands situated on the subsiding hanging-wall blocks of faults began to be occupied by volcanic rocks and their pyroclastic products were intercalated with clastic sediments. In contrast, the highlands-corresponding to the footwall blocks of faults– were undergoing uplift, emergence and erosion. These events lasted until the Callovian as indicated by the Toarcian–Bathonian gap and a new geomorphology, which was much more gentle than previously, formed by erosion of the highlands and an overfilling of the lowlands (remnant basins).

In the central and western parts of the NAPR, and sometime earlier in its eastern parts, a remarkable change occurred in the dynamics governing development of the NAPR during early Callovian–Oxfordian times. The previously eroded areas were suddenly drowned and covered by a condensed sequence of pelagic oolite- and pellet-rich shallow-marine carbonates with volcanic intercalations (Figure 2E). This condensed sequence, which is also known as an Ammonitico Rosso facies, is interpreted to be a continental plateau deposit in rift-related tectonic models (Jenkyns 1986). This sequence also indicates the presence of an oceanic realm during Callovian–Oxfordian times (Altıner *et al.* 1991). On the other hand, the sedimentation of volcaniclastic and volcanic rocks overfilled the relatively lower regions and continued with the alternation of deep-marine carbonate, current-oriented belemnite-rich tuffaceous turbiditic clastics, siliceous sediments (jasper) and spilitic pillow lavas. Thus, the whole of the NAPR was invaded by carbonate sedimentation leaving no sign of a remnant basin by Callovian–Oxfordian times (Figures 2E and 5). This regional submergence and occurrence of widespread carbonate sedimentation over all the NAPR may be ascribed to thermal subsidence taking place after crustal separation and it may correlate with transition from the syn-rift stage to the post-rift stage (Nottvedt *et al.* 1995).

In late Oxfordian–early Kimmeridgian times, the NAPR was subjected to another tectonic event recorded by new faulting (third-phase faulting), or reactivation of the faults which had resulted mainly from the second-phase faulting. This new faulting or reactivation of older faults (the inner master fault system) defined the Mudurnu, Kumluca and Akdağ basinal troughs or interior grabens (3, 8, and 12 in Figure 2E) more clearly. Thus, starting from late Oxfordian–early Kimmeridgian times, highlands corresponding to footwall blocks began to be uplifted. This event continued up to the late Valanginian and resulted in a wide carbonate platform upon which was deposited a shallow-marine regressive carbonate sequence more than 1 km thick. In contrast, during the same time interval– in the interior grabens situated on the hanging walls of faults– the subsidence and deepening continued at an increasing rate, and these grabens were filled by a sequence as much as 1.2–km–thick comprising a thinning- and fining-upward sequence of redeposited sediments (carbonate turbidites) dominated by various syn-depositional features such as thick breccias (up to 180 m thick), giant olistoliths, broken formations, and slumped deposits (up to 100 m thick). These sediments originated from the fault-bounded submarine highs (shallow-marine pelagic plateaus), and record high-rates of sedimentation and relatively rapid movement along the faults bounding the interior graben-platform system.

In the late Valanginian–early Hauterivian time interval the last and most important extensional event (fourth-phase faulting) occurred. It may have resulted from the downward pulling effect of the progressively enlarging, cooling and denser Neo-Tethyan ocean floor on the adjacent passive continental margin (Dietz 1963; Dietz & Holden 1966; Dewey & Bird 1971). This event led to intra-platform break–ups by moderate fault activity (marginal fault complex) and produced some short- and long-term intra-platform gaps (post-rift unconformity) by uplift, emergence and erosion of the fault blocks.

Finally, due to continued movement on the outer fault system, the parts of the platform adjacent to the interior graben were divided into blocks which subsided and were drowned and covered by deep-marine pelagic sediments to produce subplatforms (Figure 2E). In Aptian time and onward, the interior grabens were overfilled, while the subplatforms were onlapped by the finer basinal association. In contrast, the rest of the platforms preserved their emerged positions as palaeohighs up to the Aptian, even to Coniacian and Palaeocene times (PH in Figure 5).

A similar geologic history has been recorded from both the continental margin of the Jurassic Tethys ocean in the eastern Alps and in the Jurassic–Createaceous sequence of the northern North Sea (Eberli 1987; Nottvedt *et al.* 1995). These authors have also suggested a rift basin shaped by tilted fault blocks with depositional settings explicable in terms of the three-stage graben model. The time periods embraced by the proto-, syn- and post-rift stages of the NAPR are comparable to those of the northern North Sea rift (Nottvedt 1995) (Figure 2).

Consequently, when the substratum of the NAPR- comprising a young orogen (Şengör 1979)– is considered, the data presented here strongly support the interpretation that the NAPR constituted the south-facing passive continental margin of northern Neo-Tethys in northern Turkey and had the character of a

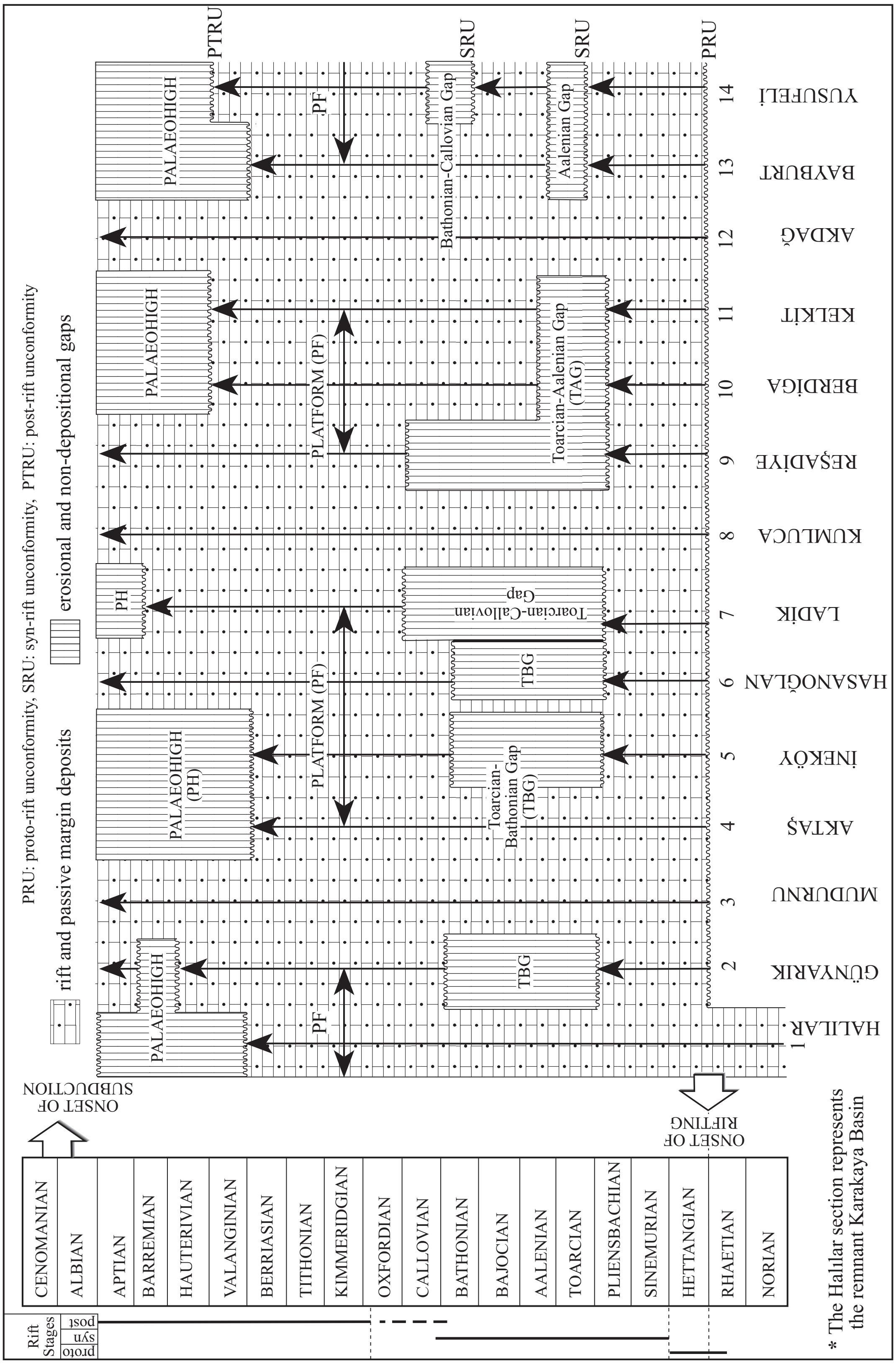

Figure 5. Diagrammatic stratigraphic columns of the north Anatolian Palaeorift showing a three-stage graben model and related unconformities.

supradetachment basin which evolved in three stages during the Hettangian–Aptian time interval (Figure 2E).

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