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Alpine tectonics of the Pannonian Basin

Tari, Gabor Csaba, Ph.D.
Rice University, 1994
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ALPINE TECTONICS OF THE PANNONIAN BASIN

by

GÁBOR TARI

A THESIS SUBMITTED
IN PARTIAL FULFILLMENT OF THE
REQUIREMENTS FOR THE DEGREE

DOCTOR OF PHILOSOPHY

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April, 1994
ABSTRACT

ALPINE TECTONICS OF THE PANNONIAN BASIN

Gábor Tari

1) Allowing for the palinspastic restoration of Tertiary strike-slip and extensional faulting it is concluded that the Pannonian Basin is superposed on a Middle Cretaceous thrust and fold belt which is an integral part of the system of the Alps, the Carpathians and the Dinarides. An upper Cretaceous flexural basin is overlying parts of that fold belt.

2) A Paleogene basin is seen as a retroarc flexural basin with respect to the Paleogene Carpathian arc.

3) The transition from an overall compressional to a transitional setting in the intra-Carpathian area occurred during the Early Miocene when large-scale transcurrent movements segmented the Alpine-Carpathian arc. Major transcurrent faults can be deduced from a study of surface and subsurface geology, but the quality of the seismic data does not permit to image them adequately.

4) The Neogene Pannonian Basin proper shows distinct modes of upper crustal extension. In the deep (>8 km) subbasins of the Pannonian Basin system like the Danube Basin the Middle Miocene syn-rift extension was accommodated by low-angle detachment faults overlying metamorphic core complexes. Other intra-Carpathian subbasins such as the Zagyva and Derecske troughs related to a transfer fault system and show moderate or negligible extension.

5) The seismic reflection profiles also suggest that the Pannonian Basin of eastern Central Europe is characterized by broad Quaternary to Recent basement upwarps that may have involved compression of the crust as a whole. These features are responsible for the outcrop distribution of the pre-Neogene "basement" of the basin. They also may suggest the beginning of a large-scale basin inversion process propagating from the W.
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CHAPTER 1
INTRODUCTION

1.1 LOCATION AND PHYSIOGRAPHIC SETTING

The Pannonian Basin is located in eastern Central Europe. It is encircled by the Carpathian Mountains to the N and E, the Dinarides to the S, and the Southern and Eastern Alps to the W (Fig. 1.1). While these mountains are characterized by relatively high average elevation (Carpathians: 1500 m; Dinarides: 1000 m; Eastern and Southern Alps: 2000 m), the Pannonian Basin s.l. is a lowland with an average elevation of 150 m above sea level. In this lowland which is about 400 km wide from N to S and 800 km long from E to W, isolated mountains (or "Inselgebirge" with a German expression) emerge from the plain with elevations up to 1000 m. These ranges subdivide the Pannonian Basin into a number of subbasins.

The Vienna Basin is not part of the Pannonian Basin s.s. It is located between the eastern termination of the Eastern Alps, the westernmost part of the West Carpathians and the Bohemian Massif (Fig. 1.2). The Danube Basin or the northwestern part of the Pannonian Basin is bounded by the Eastern Alps and the West Carpathians on the W and N, respectively, while it is bordered by the Transdanubian Central Range to the S and the E. The Great Hungarian Plain is the largest region of the Pannonian Basin; it occupies the central portion of the intra-Carpathian area (Fig. 1.2). The Transylvanian Basin located between the Apuseni Mts. and the Eastern and Southern Carpathians belongs only in a wider sense to the Pannonian Basin. All of the major basins may be subdivided into subbasins.

Two major rivers cross the intra-Carpathian region. The Danube enters the Pannonian Basin at its northwestern corner through the Vienna Basin while the Tisza, which springs from the border zone of the Western and Eastern Carpathians, enters from
Fig. 1.1. Tectonic scheme of the Alpine-Mediterranean region (Horváth, 1988).
Fig. 1.2. Main geological and geographical units of the Carpathian-Pannonian region.
the NE (Fig. 1.2). There are several lakes scattered across the Pannonian Basin. Lake Balaton, the largest of Central Europe, is located to the S of the Transdanubian Central Range. The second largest lake is Lake Fertő (Neusiedler See) at the easternmost end of the Eastern Alps (Fig. 1.2).

Together with the surrounding mountain belts, the Pannonian Basin s.l. lies within parts of nine countries: Austria, Hungary, Slovakia, Poland, Ukraine, Romania, Serbia, Croatia and Slovenia (Fig. 1.2). Such was the situation in the fall of 1993, shortly after the disintegration of the Soviet Union, Yugoslavia and Czechoslovakia. The ongoing civil war in the former Yugoslavia makes it impossible to show the state boundaries between Slovenia, Croatia and Serbia (Fig. 1.2).

1.2 PREVIOUS MODELS ON THE ALPINE EVOLUTION OF THE PANNONIAN BASIN SYSTEM

In this subchapter, previous regional models explain the tectonics of major alpine periods in the Pannonian Basin - Alps - Carpathians - Dinarides region. Unlike the rest of the thesis, here I summarize these models in chronological order. Since the earliest regional models date back to the last century a further subdivision defines contributions made before and after the application of the plate tectonic concept in the region.

1.2.1 PRE-PLATE TECTONIC MODELS

After the first contributions to the geology of Hungary (Beudant, 1822), the second half of the last century brought the first systematic geologic mapping of the isolated intra-Carpathian mountain ranges by the geologists of the Austro-Hungarian Monarchy (for a summary see Hauer, 1870). During this time the connection between the Alps and the Carpathians through the Hungarian Mid-Mountains was widely accepted (e.g. Böckh, 1872). Moreover, it became clear (Peters, 1859) that in the S, notably in the Mecsek and
Apuseni Mts. (Fig. 1.2) the Triassic and Jurassic developed in a characteristic European facies (continental "Keuper" and Liassic coal-bearing beds, respectively).

Uhlig (1903) was the first who proposed that the Mid-Hungarian Mountains form the youngest and uppermost nappe of the Carpathian system, based on its Alpine facies relationships. He also distinguished a southeastern, European-type facies block ("East Carpathian facies realm").

In contrast to this nappe-concept, Lóczy (1924) postulated a more or less rigid tectonic block in the intra-Carpathian area following an earlier conception of Mojsisovics (1880). According to Lóczy (1924): "The central zones of the Eastern Alps lose their young folds entering the Hungarian Basin and consequently continue in table-like blocks and have a block-faulted transition to the massif of the Eastern Balkan beneath the Great Hungarian Plain."

According to Kober (1923, 1931) the "Pannonian mass" is the type example of the so-called median masses ("Zwischengebirge") located in the center of young orogenes. These masses are rigid blocks and therefore in their immediate surroundings the area of orogenic folding passively follows their boundaries.

Based on this conception Prinz (1926) named this Pannonian median mass Tisia, after the Latin name of the river Tisza in Hungary (Fig. 1.3). The actual boundaries of this "ancient Pannonian mass" and their nature were subject of debate. For Prinz (1926) Tisia was the area of the Pannonian Basin and the innermost parts of the surrounding mountains (Fig. 1.3). He also visualized the rigid Tisia as having a sharp boundary with the surrounding "folded frame". According to Telegdi-Roth (1929) the boundary of Tisia coincides with the boundary of the Inner and Outer Carpathians (cf. Fig. 1.2) and this boundary is rather a transitional one with an increasing amount of folding outward. Böckh (1930) compared the Pannonian median mass to the Iranian Highland. The prevailing view of the father of the median mass concept was reiterated by his son, Lóczy Jr. (1934).
Fig. 1.3. Tisia as it was defined by Prinz (1926).
A markedly different line of thinking was represented by Pávay-Vajna (1931), who regarded the basement of the Pannonian Basin as characterized by Alpine folds and thrusts. Furthermore, Pávay-Vajna (1931) was the first who proposed that "the area of the present-day Pannonian Basin is currently in the stage of mountain-building by folding and uplift". This view was considered to be "radical" and was rejected by the geologic community at that time.

Based on his work in the Apuseni Mts., Rozlozník (1937) emphasized the geological heterogeneity of the intra-Carpathian area and pointed out that "the conception of marked contrast between the Carpathian mountain chain and the ancient Hungarian mass seems to be outdated". Szentes (1949) pointed out that the Hungarian median mass is a heterogeneous unit and the Inner Carpathian area is built up of blocks of very different origin. Moreover, the whole intra-Carpathian structural edifice was welded together and surrounded by the Outer Carpathians flysch belt only at the end of the Oligocene.

The view expressed by Horútszy (1961, 1969) also shows the alternative view of the median mass conception: "In most of the tectonic models Hungary is described as a sort of icebreaker rigid mass dividing the Alpine folds into Carpathian and Dinaric branches. Based on my experience in the Mid-mountains and the wells in the Hungarian basin, this »tectonic no man's land« displays much more Alpine and orogenic characters than we previously thought." In the works of Wein (1969), Dank and Bodzay (1971) the basement of the Pannonian Basin was described as being composed of Alpine (Cretaceous-Paleogene) overthrusts correlatable with thrust sheets in the surrounding Alpine-Carpathian-Dinaric mountain belts.

Although "Carpathian-type" rocks were encountered in many wells in the Great Hungarian Plain since the thirties, it was Körössy (1959) who introduced the term for the Szolnok flysch belt. This Cretaceous-Paleogene flysch basin in the middle of the Pannonian Basin (see Fig. 2.4) raised serious doubts about the homogeneity of the Tisia
block. Based on geophysical arguments Szénás (1968) rejected the classification of the Szolnok succession as a flysch basin, and it was reclassified by some as only a "flyschoid" basin (Juhász, 1968). However, following Szepesházy (1973), the Szolnok flysch basin was correlated in an ill-defined way with the Carpathian flysch belt by most authors (e.g. Körössy, 1977; Balla, 1982; Nagyamarosy, 1990a), indicating large strike-slip movements within the intra-Carpathian region.

In the sixties, the median mass concept was still accepted by many with only small modifications (e.g. Schmidt-Eligius, 1947; Vadász, 1955, 1961; Slavin, 1961; Szalai, 1961, 1970). The map and cross-section of Schmidt-Eligius (1961) shown in Fig. 1.4 clearly reflect the long-lasting influence of the "Zwischengebirge" concept of Kober (1931).

At the same time, new geophysical methods made an important impact. Gálfi and Stegena (1960) were the first to determine the crustal structure of the Pannonian Basin by deep seismic soundings. As it was further confirmed by the measurements of Mituch (1964), the crust beneath the Pannonian Basin turned out to be significantly thinner (24.5-29.5 km) than the world average. The magnetotelluric work of Ádám (1964) and the geothermic research of Stegena (1964) also supported the thinned and therefore "hot" character of the crust in the intra-Carpathian area.

Based on these new geophysical findings and adopting Van Bemmelen's (1966) ideas, Szádeczky-Kardoss (1967) introduced the so-called "geotumor" concept. In this model the passive median mass was replaced by an active mantle diapir beneath the Pannonian Basin causing high heat flow and thin continental crust. Due to uplift in the center gravitational nappes glided outwards, first forming the Cretaceous Inner Carpathian nappes and then the Tertiary flysch belt. Later the uplifted central part collapsed, creating the present-day Pannonian Basin, surrounded by thrust-fold belts.
Fig. 1.4. The map and cross-sectional view of Tisia by Schmidt-Eligius (1961).
1.2.2 PLATE TECTONIC MODELS

After more than a century, the original observations of Peters (1859) on the contrasting Early Mesozoic facies patterns in the Pannonian Basin were reconfirmed by Géczy (1973). This led to the redefinition of Tisia as the southeastern half of the intra-Carpathian region (Fig. 1.5). This new Tisia (or Tisza) includes the area of the Mecsek, Villány, Papuk and Apuseni Mts. (see Kovács, 1982, for a summary). Géczy (1973) was the first who pointed out that this newly defined Tisia is located farther to the S than it should be, i.e. on the northern, European margin of the Tethys based on its characteristic European facies development. Moreover, just to the N of Tisia, the Northern Pannonian unit (or Pelso unit after the Latin name of Lake Balaton) has a facies characteristic for the southern, African margin of the Tethys. Recently, Kovács et al. (1989) compiled the early Mesozoic facies zones of all the major tectonic units in the Pannonian Basin - Alps - Carpathians - Dinarides region shown in approximated present-day coordinates (Fig. 1.6). This greatly simplified, but useful scheme shows the proximal or distal position of different subunits within the major blocks in relation to the Tethyan continental margin and the oceanic basin. The most important element in this picture is the "exotic" position of the South Pannonian or Tisia block.

There is an ongoing debate about the European origin of Tisia (see Misik and Kázmér, 1989, for an overview), an interpretation that is not accepted by a number of Slovakian workers (e.g. Misik et al., 1989). The European origin of Tisia, however, is the key to the reconstruction of the intra-Carpathian area (see Chapter 6). If the "exotic" position of Tisia is accepted, large-scale strike-slip movements must be responsible for the present-day situation. Laubscher (1971) reached the same conclusion on a more regional scale.

For the North Pannonian or Pelso block, Majoros (1980) made a regional compilation on the distribution of Upper Permian facies. He found that a N-S trending
Fig. 1.5. Position of Tisia in the Alpine-Carpathian-Dinaric system (Kovács et al., 1989).
Fig. 1.6. Schematic distribution of early Mesozoic facies zones in Tisia and the adjacent eocalpine tectonic units in the Alpine-Carpathian-Dinaric system (Kovács et al., 1989).
continental/marine facies boundary to the E of Lake Balaton in the Pannonian Basin is correlatable to the same facies boundary in the Southern Alps. This suggests an approximate 400-500 km right-lateral offset along the Periadriatic Line (Majoros, 1980). Kázmér (1984) also suggested that strike-slip movements of about 450-500 km amplitude could explain the observed mismatches of Permian-Mesozoic facies boundaries. Kázmér and Kovács (1985) published a series of Mesozoic reconstructions to illustrate these facies mismatches. According to their model, the bulk of the northern part of the Pannonian Basin was squeezed out from the Alpine realm between the southern right-lateral Periadriatic - Balaton Line and the northern left-lateral DAV - Rába - Hurbanovo - Diósjenő Line (Fig. 1.7). Since Kázmér and Kovács (1985) found mismatches not only in the Paleozoic-Mesozoic succession but also in the Eocene facies distribution, they proposed a Middle Eocene - Late Oligocene age for the continental escape (sensu Burke and Sengör, 1986) of the Northern Pannonian unit from the Alpine realm.

Most of the models that address the problem of Paleogene tectonics in the Pannonian Basin are related to this continental escape model. For this time interval which spans about 30 Ma, Kázmér (1984), Fodor and Kázmér (1989), Fodor et al. (1992) assumed a rather continuous right-lateral slip along the southern margin of the escaping North Pannonian block, creating the pull-apart basin complex of the Hungarian Paleogene Basin (Fig. 1.8a).

A slightly different line of thinking is followed in Báldi and Báldi-Beke (1985) and Royden and Báldi (1988). According to these authors the Hungarian Paleogene Basin was formed in a broad continental transform zone connecting areas of active shortening in the outer Carpathians to areas of contemporaneous shortening in the Dinarides (Fig. 1.8b). In this model the sense of shear in the ENE-trending transcurrent zone was also dextral, such as in the models of continental escape. An alternative model which I proposed (Tari, 1992a; Tari et al., 1993) is shown in Fig. 1.8c and will be discussed in Chapter 5.
Fig. 1.7. Cartoon showing the exotic position of the Drauzug and the Transdanubian Central Range (adapted by Balla, 1988, from Kázmér and Kovács, 1985). A, Present situation; B, Restored situation. DAV, Defereggental-Anterselva-Valles Line.
Fig. 1.8. Cartoon showing the contrasting geodynamic models on the formation of the Hungarian Paleogene Basin (HPB). All the models show a Late Eocene reconstruction roughly following the palinspastics proposed by Pescatore and Slaczka (1984) and Csontos et al. (1992). The present-day outline of the Carpathian front is indicated with a dashed line. a) The HPB formed as a pull-apart basin complex related to continental escape (e.g. Kázmér, 1984). b) The HPB evolved in an intercontinental transform fault zone as a series of pull-apart basins (e.g. Royden and Báldi, 1988). c) The HPB was a flexural basin in the back of the compressional Western Carpathians (this work).
As to the Neogene Pannonian Basin, the concept of plate tectonics was introduced by Szádecky-Kardoss (1971). Unfortunately, this first attempt which outlined only three subduction zones in this region was shortly followed by extreme models, e.g. Szádecky-Kardoss (1975) postulated eight(!) subduction zones in the very same region.

Stegena (1972), Horváth (1974), Horváth et al. (1974) classified the Pannonian Basin as a Mediterranean interarc basin, where the subduction-generated mantle diapir is responsible for the thinning of the crust by subcrustal erosion. Channel and Horváth (1976) proposed a plate tectonic model for the Mesozoic-Cenozoic evolution of the Central Mediterranean, introducing the Adriatic microplate, as the promontory of Africa.

Bally and Snelson (1980) classified the Neogene Pannonian Basin as a Mediterranean back-arc basin characterized by continental crust, where extension did not advance until opening of an oceanic basin, in contrast to W-Mediterranean basins. Horváth and Berckhemer (1981) also suggested that the Mediterranean back-arc basins actually represent different evolutionary stages in back-arc extension (see Fig. 1.1).

The Neogene Pannonian Basin was one of the first areas where McKenzie's (1978) pure shear model of lithospheric extension was tested (Sclater et al., 1980). Analysis of subsidence curves and thermal data suggested that a modified lithospheric extensional model (Royden et al., 1983a), whereby large amounts of heat are added to the uppermost mantle during extension, produced good agreement with the observed heat flow, thermal gradients, rates of thermal subsidence and vitrinite reflectance. In Royden's model, the initial crustal thinning phase occurred during the Middle Miocene (17.5-10.5 Ma), while the subsequent late Miocene to Recent period (10.5-0 Ma) corresponds to the thermal subsidence phase. Crustal extension beneath most parts of the basin is estimated to have been about 50% to 120% (Horváth et al., 1988).

Horváth and Royden (1981), Royden et al. (1982) recognized the importance of strike-slip faults in the opening of discrete basins in the Pannonian area. In their model the
extension occurred along a conjugate system of strike-slip faults (Fig. 1.9) that connected
areas of coeval extension to one another and also to areas of coeval shortening in the
surrounding fold-and-thrust belt in the Carpathians. Since the Hungarian part of the
Pannonian Basin system is mainly characterized by a thick Neogene to Quaternary
sedimentary fill, the structure of the individual basins can be studied only by reflection
seismic and drillhole data. The map shown in Fig. 1.9 is largely based on the interpretation
of subsurface data by Rumpler and Horváth (1988).

Royden et al. (1983a) and Royden (1988) made the point that the extension in the
Pannonian Basin and coeval compression in the Carpathians were the result of the
continued subduction of the European plate, probably driven by negative buoyancy of the
subducted slab. In this model back-arc extension occurs to fill the gap created by retreat of
the European slab due to roll-back of the subducted slab (see also Royden, 1993). More
recently Royden and Burchfiel (1989) explained the extension of the Pannonian Basin in
terms of plate convergence and subduction rate.

In a recent synthesis on the structure of the Eastern Alps, Ratschbacher et al. (1991)
suggested extrusion tectonics as the driving mechanism for the Neogene extension in the
Eastern Alps and intra-Carpathian region. Extrusion tectonics is defined as a synchronous
interaction between tectonic escape (Burke and Sengör, 1986) and extensional collapse
(Dewey, 1988). Ratschbacher et al. (1991) concluded that the forces applied to the
boundaries of the Eastern Alps, causing their continental escape, seem to be of similar
importance as the extensional spreading of gravitationally unstable crust. This argument
will be further refined in Chapter 8, by discussing the timing of the continental escape
episode and the back-arc extension as it is reflected in the structure and stratigraphy of the
Pannonian region. Note that Ratschbacher et al. (1991) placed the continental escape in
the Miocene in contrast to others (see above).

Doglioni et al. (1991) and Doglioni (1992) also proposed a model of Miocene back-
Fig. 1.9. Tectonic map of the Pannonian Basin and surrounding regions showing the main faults and folds of Neogene age (Rumpler and Horváth, 1988). Legend: 1, Molasse foredeep; 2, Alpine-Carpathian flysch belt; 3a, Inner Alpine-Carpathian mountain belt and the Dinarides; 3b, outcrops of Neogene calcalkaline volcanic rocks; 4, strike-slip faults; the sense (and usually the amount) of displacement is well constrained (thick arrows) or unconstrained (thin arrows); 5, normal fault, thrust fault and fold; 6, areas of major crustal extension and subsidence.
arc extension driven by subduction roll-back to explain the formation of the Pannonian Basin. This roll-back of the west-dipping European continental margin, according to Doglioni (1990), is the consequence of an eastward directed mantle flow.

Balla (1984) attempted to explain the formation of the Neogene Carpathian loop using a map-view kinematic approach. In his model (Balla, 1984; 1987, 1988, 1990) he subdivided the present-day intra-Carpathian area into two major blocks, roughly equivalent of the Pelso and Tisia blocks of others (Fig. 1.10a). The model is essentially based on anomalous paleomagnetic declinations (~100° clockwise) measured in the Mecsek Mts. of the southern block. Assuming rigid body rotation, Balla (1984) rotated back the southern unit to the S to arrive at the restored palinspastic scheme of the Early Miocene (Fig. 1.10b).

Recently, Csontos et al. (1992) also proposed a map-view reconstruction of the Pannonian Basin area. This reconstruction is based on the restored position of the Tertiary Outer Western Carpathians nappes (Oscypko in Csontos et al., 1992). As a first-order kinematic analysis this model is accepted in this thesis and it serves as the starting point for a number of new plate-tectonic reconstructions (see Chapter 6).

1.3 SCOPE OF STUDY

About 90% of the pre-Tertiary "basement" and about 60% of the pre-middle Miocene strata are covered by the sedimentary fill of the Pannonian Basin s.s. This clearly shows the importance of subsurface geology. Many of the above described geological problems in the Pannonian Basin, debated for almost a century, may be resolved only by the systematic evaluation of all subsurface geologic information, most importantly from the interpretation of seismic data combined with well information. The "classical" problems of Hungarian geology that will be addressed in this thesis follow:
Fig. 1.10a. Tectonic subdivision of the northern part of the Central Mediterranean in the present-day situation (Balli, 1986). For letter symbols see Fig. 1.10b on next page.
Fig. 1.10b. Structural reconstruction of the Central Mediterranean 20 Ma before (Balli, 1986). For legend see Fig. 1.10a on previous page.
1) The existence of Cretaceous Alpine nappes in the pre-Tertiary basement of the Pannonian Basin has been debated since the beginning of the century. Although in many boreholes repeated sections were found, many were interpreted as local disturbances along major fault zones in the Pannonian Basin. The large-scale allochthoneity of the Transdanubian Central Range was recently suggested in contradiction to the more traditional view, that claims negligible Alpine deformation in this major tectonic unit. A closely related problem is the much-debated continuation of the magnificent nappe pile of the Eastern Alps underneath the Danube Basin towards the Transdanubian Central Range and the Western Carpathians.

2) The Upper Cretaceous sedimentary succession beneath the Pannonian Basin probably developed in a different sedimentary facies from the well-known "Gosau" basins of the Eastern Alps. These basin fragments are considered post-tectonic with respect to the major overthrowing events in the alpine realm. Based on their facies development, these basins are classified as "epicontinental", a rather meaningless descriptive term.

3) Similarly, the enigmatic Paleogene basin segments beneath the intra-Carpathian area are poorly understood. These basins are traditionally classified either as "flysch" or as "epicontinental", without referring to a geodynamic context. Recently a pull-apart origin was proposed by analogy to the overlying, much better understood Pannonian Basin s.s. Such a backward extrapolation of a transtensional model - as will be shown later - does not seem tenable for a number of reasons and requires an alternative geodynamic scenario.

4) For the last decade, the Paleogene succession has also been considered as the manifestation of a major "continental escape" event, that has significantly reshaped the paleogeography of the Alps-Carpathians-Pannonian Basin system. Very different models were proposed involving the exact timing, magnitude and geometry of this escape (or "extrusion" of some authors). The restoration of the eastwardly escaping intra-Carpathian tectonic units to their original place within the Alpine edifice is also a subject of major
controversy. The most common approach involves correlating Mesozoic isopic facies zones, but more reliable and accurate structural markers ought to be used for better restorations.

5) The efforts of scientists during the last decade certainly improved our understanding of the Neogene evolution of the Pannonian Basin s.s., but some fundamental questions remain unanswered. The Neogene Pannonian back-arc basin was superimposed on an earlier Alpine compressional realm. Thus the compressionally pre-conditioned "memory" of the pre-Neogene basement supposedly influenced the magnitude and geometry of the subsequent continental extension by reactivation of regional decollement levels. Similarly, the transfer role of many of the Neogene strike-slip(?) faults connecting areas characterized by extension of very different magnitude has always been assumed without adequate documentation. The amount of extension in a given subbasin has been estimated on the basis of subsidence analysis only, without extensive use of regional reflection seismic data.

6) Quaternary tectonics are commonly overlooked in the Pannonian Basin. The very recent and, in some cases, currently active fault zones showing typical flower structures on seismic sections are erroneously assigned to the Neogene formation of the Pannonian Basin. These neotectonic features and the local broad upwarping of the basement from below the Neogene succession are responsible for the characteristic "Inselberg"-pattern of present-day outcrops. The presently active tectonics are related to a new stage in the Alpine evolution of the Pannonian Basin.

1.4 DATA BASE

Abundant subsurface data are available from most of the Pannonian Basin. Within Hungary alone, more than 11,000 km of wildcat and production wells have been drilled, and more than 35,000 km of multifold stacked seismic profiles exist.
This study is primarily based on exploration reflection seismic data, acquired and processed by the Hungarian Geophysical Exploration Company, Budapest. In the construction of regional profiles crossing the Pannonian Basin of Hungary, I selected some 2500 km of reflection seismic data. For a detailed analysis of the actual study area, which is located in the Hungarian part of the Danube Basin (Fig. 1.2), I interpreted about 150 reflection seismic sections over about 4000 km. Although this data set is composed of lines from different vintages, generated with different sources (explosion and Vibroseis), all of them are 24-fold migrated sections. Most commonly the seismic information was processed to 4 s TWT time.

This data set was supplemented by eight crustal seismic reflection lines in the actual study area, acquired and processed by the Lóránd Eötvös Geophysical Institute of Hungary, Budapest. These lines were processed to 10 s TWT time.

The interpretation of the regional profiles was aided by the general description of some 120 wells. In the study area of Western Hungary, I used an additional 450 wells, mostly drilled by the Hungarian Oil Company (MOL). Detailed well-log information on about 60 selected wells was also analyzed and integrated into the present study. Velocity information and time-depth functions were provided for some 20 key boreholes in the area.

A large number of published geological maps were also utilized for better constraints by outcrop information into subsurface projections. A list of these maps, with scales ranging from 1:20,000 to 1,000,000, can be found in the Appendix. Measured sections of outcrops were collected and analyzed largely from published material and supplemented by a limited amount of my own work, carried out in the summers of 1990, 1991 and 1992.

1.5 ORGANIZATION OF THIS STUDY

The next eight chapters are organized in the following manner. In Chapter 2, a broad
stratigraphic and tectonic outline of the Pannonian Basin and the surrounding thrust-fold belts is presented. Many of these areas are considered classical sites; consequently an enormous number of scientific papers dealing with the geology of this region have been published during the last one and a half centuries. The main aim of Chapter 2, however, is to give an overview on the present-day geologic understanding of the broader Alpine framework for this study without entering into the specifics. Oversimplifications in this chapter were unavoidable.

Chapter 3 offers information rather than models for the central and Hungarian portion of the Alps-Carpathians-Dinarides-Pannonian Basin system (Fig. 1.2). For the first time, regional seismic sections were constructed across the whole country, combining industry and academic reflection seismic data. To give a unified picture of this data set, three seismic panels show the 18 regional seismic sections as line drawings. Although this presentation reflects my own interpretation, it is still a relatively objective way to legibly display the geology of the Pannonian Basin. Chapter 3 is also supplemented with selected published seismic sections and their interpretations to put the regional sections in a more complete perspective.

Chapter 4 discusses the stratigraphy of the NW part of the Pannonian Basin (Fig. 1.2). In contrast to Chapter 2, this chapter summarizes not only the Alpine (Permian-Recent) stratigraphy but also includes the description of the pre-Alpine, Hercynian and Caledonian basement of Paleozoic age. The stratigraphic overview in this chapter gives specifics on the scale of individual formations. This chapter also summarizes the structural models proposed so far for the pre-Alpine and Alpine structural evolution of the study area.

Chapter 5 presents the data and interpretation for the study area of NW Hungary. All of the original reflection seismic data in this chapter are accompanied by their line drawing interpretations, but a significant number of seismic lines were reproduced only as
line drawings because of the relatively poor data quality. The seismic reflection and well data allow specific problems of the Alpine tectonic evolution to be addressed. The seismic data, however, do not permit adequate study of certain details of the Alpine structural evolution. Therefore, the tectonic history of these later stages is discussed on the basis of well data and field relations. I tried to separate interpretations supported by data from models which were only inferred from the incomplete data set. The chapter is organized according to the main stages of Alpine tectonics.

Chapter 6 returns to the more regional problems of the Alps-Carpathians-Dinarides-Pannonian Basin system, on the basis of the results described in Chapter 5. Since the study area lies in a key area of the Alps-Carpathians transitional zone, many findings have important consequences for the broader tectonic picture of Central Europe. These tectonic and geodynamic extrapolations are supplemented with new palinspastic reconstructions for the Alpine history.

Chapter 7 briefly outlines some interesting ancient and recent worldwide geologic examples (basins and thrust-fold belts) which may serve as a useful analogue for certain periods in the Alpine evolution of the Pannonian Basin.

In Chapter 8 some of the still debatable problems are discussed and possible explanations are offered without detailed documentation.

Chapter 9 briefly summarizes the original observations and models of this study as a conclusion.

The Appendices include supplementary data and interpretations that are not closely related to the main subject of this thesis.
CHAPTER 2

ALPINE (PERMIAN-RECENT) STRATIGRAPHIC AND TECTONIC OUTLINE
OF THE PANNONIAN BASIN AND THE SURROUNDING MOUNTAIN BELTS

2.1 OUTLINE OF MAJOR TECTONOSTRATIGRAPHIC UNITS

The Pannonian Basin occupies the central part of the European Alpine belt (Fig. 2.1). At the western margin of the Pannonian Basin, the thrust-fold belts of the Western, Eastern and Southern Alps branch to the Dinarides to the SE and to the Western Carpathians to the NE. While the Dinarides continue to the SE with only a minor change in strike into the Hellenides, the Western, Eastern and Southern Carpathians form an almost complete loop before continuing into the Balkans. Thus the Alpine chain is about 300, 1000 and 400 km wide in the Alps, Dinarides-Pannonian Basin-Carpathians and in the Hellenides-Balkans, respectively. The pronounced widening in the Pannonian sector is mainly the result of the Neogene extensional basin formation in this central area. The surrounding Eastern Alps, Carpathians and Dinarides indeed project below the Pannonian Basin, which is superimposed on these Alpine thrust-fold belts.

The present-day assembly of the superimposed basin complex and the underlying structures of very different style and age in the Pannonian Basin-Alps-Carpathians-Dinarides system suggest a very complicated origin. Individual tectonostratigraphic units can be clearly outlined for distinct Alpine orogenic regimes, but any subdivision is valid only for very limited time periods (see below). The major periods of a geodynamic scenario for the Alps-Carpathians-Dinarides-Pannonian Basin system are:

a) Late Permian - Late Jurassic. Two periods of continental breakup occurred during this time. While the Triassic rifting aborted in most of the area, the Jurassic extension led to the formation of the Tethys ocean.

b) Early Cretaceous - Paleocene. This was a time of major overthrusting in the Alpine
Fig. 2.1. Tectonic sketch map showing the position of the Carpathian Mountains and the Pannonian Basin within the Alpine belts of eastern Central Europe (Royden, 1988). Stippled area indicates parts of the Pannonian Basin where the depth to base of Miocene exceeds 3 km. Subbasins: V, Vienna; D, Danube or Little Hungarian Plain; G, Graz; Z, Zala; Dr, Dráva; S, Sava; Tc, Transcarpathian; Ts, Transylvanian; GHP, Great Hungarian Plain; B, Banat. Other abbreviations: P.A.L., Periadiatric Line; P.K., Pieniny Klippen Belt; M.M., Hungarian Mid- (or Central) Mountains.
system, when many of the oceanic troughs and passive margins disappeared. This interval is traditionally subdivided into several periods of compressional events (see Chapter 4).  
c) Eocene - Oligocene. This interval represents the second major period of compression and collision in the Alpine thrust-fold belt, following a period of relative tectonic quiescence. This period shows many similarities with the former one.  
d) Miocene - Recent. While continuing convergence between Europe and Africa still forms thrust-fold belts in certain areas, in the Pannonian Basin this interval is largely characterized by the opening of superimposed extensional basins. The marked change from the previous stage was due to a large-scale continental escape period in the Early Miocene which disintegrated the former Alpine thrust-fold belt.  

These broad periods are close to those defined by Trümpy (1973) in the classical Swiss sector of the Alps. Although there are slight differences in the timing (cf. also Trümpy, 1980), I will refer to his terms for the major Alpine stages outlined above. These are the early Alpine, the Paleoalpine (or Eoalpine), the Mesoalpine and the Neoalpine stages, respectively.  

The structure of the mountain belts surrounding the Pannonian Basin was formed during the above Alpine stages. All these belts have a dominant vergence outward from the intra-Carpathian basins (Fig. 2.1). Within the thrust-fold belts, an internal Eoalpine zone of Mesozoic nappes is distinguished from an outer Tertiary belt of Eoalpine and Neoalpine nappes. The internal zone is somewhat discontinuous, but the outer flysch belt is uninterrupted at the perimeter of the Alps and the Carpathians (e.g. see Fig. 1.2).  

In the intra-Carpathian area, the main Alpine stages outline three major tectonostratigraphic levels. An idealized geologic section in the center of the basin illustrates these major units (Fig. 2.2). The upper, Neoalpine level comprises the Neogene Pannonian Basin s.s. The middle, Mesoalpine level consists of Paleogene basins. The underlying basement level is a complicated assembly of Mesozoic successions formed
Fig. 2.2. Idealized scheme of major tectonostratigraphic levels of the Pannonian Basin.
during the early and Eoalpine stages.

In map view the Neoalpine level is best characterized by its isopach map (Fig. 2.3). The Miocene-Quaternary succession covers almost the entire area and locally is very thick (>8 km), outlining a number of subbasins. Neogene volcanics outcrop at the basin margins but they are also abundant in the subsurface (see Fig. 2.2). In contrast to the Neoalpine level, the underlying Mesoalpine level forms isolated basins with different lithofacies (Fig. 2.4). This map shows the surface and subsurface distribution of the Paleogene mostly underlying the Neoalpine level (see Fig. 2.2). The numerous tectonostratigraphic units of the lower level (Fig. 2.5) consisting Mesozoic-Paleozoic rocks are distinguished by their early and Eoalpine lithofacies and/or structure. The present-day collage of these units developed during the subsequent Mesoalpine and Neoalpine periods.

The following subdivision of the Alps-Carpathians-Dinarides-Pannonian Basin system mainly follows the units underlying the superposed basins. Eight major tectonostratigraphic provinces are shown in Fig. 2.6 with further internal subdivisions summarized in stratigraphic tables (Figs. 2.9, 2.11, 2.13, 2.15, 2.17, 2.20, 2.22; the legend of lithostratigraphic symbols used in this thesis is shown in Fig. 2.7). This simplified scheme is only one of many (e.g. Mahel, 1974; Balla, 1984, 1987a; Földváry, 1988; Horváth and Galácz, 1991, Csontos et al., 1987; Csontos and Vörös, in press). However, the subdivision adopted here emphasizes the similarities between certain subunits which can be further used as broad markers for the palinspastic reconstruction of the present-day Pannonian Basin-Alps-Carpathians-Dinarides system. The geographic boundaries of these major units and their internal subunits are defined in the introductory part of subchapters 2.2 and 2.3.

2.2 ALPS-CARPATHIANS-DINARIDES THRUST-FOLD BELTS

The general stratigraphy and tectonic evolution of the thrust-fold belts surrounding
Fig. 2.3. Miocene-Quaternary isopach map of the Pannonian Basin (Royden and Dövényi, 1988). Isopach lines for every kilometer interval greater than 3 km are indicated by white lines on black background. Approximate ages of igneous rocks are shown. Subbasins: V, Vienna; D, Danube or Little Hungarian Plain; G, Graz or Styrian; Z, Zala; Dr, Dráva; S, Sava; Tc, Transcarpathian or East Slovakian; Ts, Transylvanian; P, Great Hungarian Plain.
Paleogene Basins of the Intra-Carpathian Region

Fig. 2.4. Distribution of Paleogene basins in the intra-Carpathian region.
Fig. 2.6. Index map of simplified tectonic maps of major tectonostratigraphic units.
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Fig. 2.7. Legend of lithostratigraphic columns of this thesis.
the Pannonian Basin are described below beginning with the Eastern Alps. Then, following a clockwise direction the Western, Eastern and Southern Carpathians, the Dinarides and the Southern Alps are discussed. For simplicity, that part of the European foreland located in front of the above mountain belts is included.

2.2.1 EASTERN ALPS

The following very brief summary of the geology of the Eastern Alps is based largely on the synoptic works of Gwinner (1971); Janoschek and Matura (1980); Oberhauser (1980); Tollmann (1977, 1986); Flügel and Faupl (1987) and Neubauer (1992). Appendix A gives more details on the stratigraphy and tectonic evolution of the eastern end of the Eastern Alps.

The major Neogene transcurrent fault of the Periadriatic (or Insubric) Line to the S forms the sharp southern boundary of the Eastern Alps (Fig. 2.8). To the W, the boundary is drawn near the Swiss-Austrian border where the Austroalpine nappe units overthrust the Helvetic complex. To the N, the Molasse Basin can be regarded as a geological border of the Eastern Alps, while to the E, the transition to the Western Carpathians lies beneath the Vienna and Danube Basins.

2.2.1.1 STRATIGRAPHY OF THE EASTERN ALPS

The Alpine stratigraphy of the N-vergent Eastern Alps is summarized below from N to S (structurally from the lowest to the highest units, Fig. 2.9). The European foreland outcrops to the N in the Bohemian Massif, which projects with its Mesozoic cover below the Eastern Alpine thrust-fold belt. The Mesozoic succession below the Oligocene-Miocene molasse has a typical Germanic neritic facies (Swabian), extending from the Triassic to the Lower Cretaceous. The upper Cretaceous succession shows pelagic influence reflecting the load of the Alpine thrust sheets to the S. This Mesozoic succession
Fig. 2.9. Alpine stratigraphy of the Eastern Alps. The location of different units is shown in Fig. 2.8a.
is missing on the basement promontory of the Bohemian Massif (Wessely, 1987).

The Molasse Basin is overthrust by the Helvetic and Ultrahelvetic nappes, a foreland thrust-fold belt. These nappes are further subdivided into the northerly Helvetic s.s. and the southerly Ultrahelvetic unit. This subdivision is based on lithological differences, since the Helvetic subzone is dominated by shallow-water carbonates, whereas the Ultrahelvetic is characterized mainly by a marly facies. This difference reflects their relative paleogeographic position on the European passive margin with respect to the deep basin to the S. In the eastern part of the Eastern Alps the Ultrahelvetic subzone is called the Gresten zone with its characteristic coal-bearing Lower Jurassic facies.

The Rhenodanubian Flysch units were deposited in front of the Alpine edifice during Cretaceous-Eocene orogeny. The Barremian onset of flysch deposition coincided with the beginning of the southward subduction of Europe in the Hauterivian-Barremian in the Eastern Alps (e.g. Tollmann, 1987).

The Penninic unit is made up of strongly metamorphosed "lower plate" successions stacked as a number of internal nappes that outcrop as tectonic windows (Engadin, Tauern, Rechnitz) underneath the "upper plate" Austro-Alpine nappe complex (Fig. 2.8). The southern part of the originally long, narrow continental fragment of the Middle Penninic is preserved in the Centralgneiss nappes of the Tauern window. This thick mass of Hercynian granitoids has a thin Mesozoic cover (Hochstegen facies) metamorphosed in greenschist facies. The succession of the Jurassic South Penninic oceanic basin is found in the Schieferhülle nappe pile of the Tauern window (Frisch et al., 1987; Höck and Koller, 1989; Koller and Höck, 1990). In the Rechnitz window the South Penninic sequence seems more complete (e.g. Pahr, 1980) and in a calciphyllite unit Cretaceous sponge spicules were found (Schönlau, 1973).

The Austroalpine nappe complex dominates the Eastern Alps and consists of basement units and their sporadic Permo-Mesozoic cover in the Central Eastern Alps (i.e.
the area between the Periadriatic Line and the Graywacke Zone), while the Northern
Calcareous Alps are characterized by mainly Permo-Mesozoic units that are detached from
their original basement (Fig. 2.8). The Austroalpine nappes also cover the Rhenodanubian
flysch, the Helvetic nappes, and the downdip extension of the Molasse Basin.

The Triassic succession in the Lower Austroalpine is thicker than the Penninic
Triassic. The Mesozoic greenschists of the Lower Austroalpine of the Radstatt Tauern are
similar to the Jurassic of the adjacent Penninic zone (schists and breccias). The thick
Jurassic mass-flow breccias (Hochfeind facies) are related to steep fault scarps, perhaps
indicating fault activity (Häusler, 1987). The facies of the Upper Triassic changes from the
W (Radstatter Tauern) to the E (Semmering) from Hauptdolomite (peritidal carbonate
facies) to Carpathian Keuper (continental variegated clays). The main characteristics of the
basement nappe of the Middle Austroalpine is its areally very limited Mesozoic cover
(Stangalm Triassic).

In contrast, the Upper Austroalpine nappe complex is characterized by the
dominance of the Permo-Mesozoic cover. Even though the Upper Austroalpine nappe
system once covered the entire Eastern Alpine realm, today its main mass is located in the
Northern Calcareous Alps, where it comprises several thrust sheets. From bottom-to-top
these thrust units are the Bajuvaric, the Tirolic and the Juvavic, with the involvement of
deeper stratigraphic levels in the higher units. These nappes have a distinct Upper Triassic
lagoonal and reefal platform to basinal carbonate facies (Hauptdolomit, Dachstein
Limestone and Hallstatt Limestone, respectively). The southern border of the Northern
Calcareous Alps is the Graywacke Zone consisting of anchimetamorphic Paleozoic rocks.

The southernmost part of the Upper Austroalpine unit is exposed just N of the
Periadriatic Line, in the Drauzug area. This region, referred to as Licitum by Tollmann
(1987), has an Upper Triassic facies different from that of the Northern Calcareous Alps
(e.g. Bechstädt, 1978). The Lavant flysch was deposited during Albian times.
2.2.1.2 TECTONIC EVOLUTION OF THE EASTERN ALPS

The detailed early Alpine history of the Eastern Alps described here applies also with certain modifications to other thrust-fold belts described later.

After the Hercynian orogeny, the area of the original Eastern Alps became uplifted and dominated by continental sedimentation. During Late Permian, the sea slowly invaded this region from the SE (e.g. Laubscher and Bernoulli, 1977). The Lower Triassic neritic sandy shales and sandstones were generally replaced in the Anisian by carbonate platforms which produced very thick (up to 2-3 km) sedimentary successions due to accelerated subsidence. Deep, apparently fault-controlled troughs became rather extensive by the end of the Anisian. The distribution of the corresponding deep-water Hallstatt facies indicates that the deep troughs were located generally to the S of the broad North Alpine carbonate platform area (e.g. Lein, 1987). These basins display the main characteristics of an initial rifting phase, which was "aborted" later (e.g. Bechstädt, 1978).

During the Carnian, relief rejuvenation due to tectonics of uncertain origin in the continental hinterland resulted in increasing terrigenous influx to the shelf area. This terrigenous material transported from the N was trapped in the wide Raibl Basin. Interestingly enough, the platform to the S of this region shows no signs of this event.

The Late Triassic as a whole is characterized by the formation of a broad carbonate platform which compensated the relief differences of the Middle Triassic. This reflects a thermal subsidence phase perhaps following the previous rifting episode. This pattern was interrupted during the Late Norian, between the deposition of the Hauptdolomite and the Dachstein Limestone, when the deeper water Kössen Basin was formed in the area of the earlier Raibl Basin, again suggesting strong terrigenous influx. Certain parts of the southern platform region drowned during the Late Triassic merging of the epipelagic Hallstatt facies region.

In the Early Jurassic the centers of maximum subsidence shifted to the northernmost
part of the North Alpine shelf as the result of the initial extension of the South Penninic Ocean. In contrast to the Middle Triassic rifting, this extension eventually resulted in the opening of two oceanic troughs (North and South Penninic) to the N of the North Alpine carbonate realm. By the end of the Jurassic, deep-water environments with radiolarite deposition prevailed in the whole region.

In the southern part (Juvavicum) of the Eastern Alpine section gravity tectonics occurred during the Malm and the early Neocomian (Tollmann, 1987b). During this interval, the initial shortening reactivated the former normal fault planes as thrusts and uplifted certain basement blocks, separating the sedimentary cover from its substratum. The resulting gravity nappes (e.g. Berchtesgaden, Hallstatt, Dachstein, Schneeberg and Mürzalpen nappes) glided towards the N over long distances.

Due to the N-directed Eoalpine events previously described, roughly E-W trending facies zones are stacked in the Alpine nappe pile, with the southernmost zone in the uppermost position. The Alpine orogeny started in the Early Cretaceous. During the Neocomian, in the area of the later Calcareous Alps, an elongated, deep sea trough was formed, with typical flysch and mass-flow deposits (Rossfeld facies). These sediments were deposited in a trench in front of a N-vergent accretionary complex (e.g. Decker et al., 1987). Indeed, the earliest radiometric ages of Alpine metamorphism range between 135-130 Ma (Kralik et al., 1987). Thus the Hauterivian-Barremian time interval represents the first phase of Eoalpine orogeny, when subduction began in the Austroalpine and Penninic domains. During the Albian-Turonian, the Austroalpine nappe pile overthrust the South and Middle Penninic facies zones.

Deposition of the Senonian Gosau facies began in the Coniacian, and it is regarded as a post-tectonic succession commonly overlying Eoalpine thrust contacts. The Gosau basins of the Northern Calcareous Alps are typically involved in later, Mesoalpine thrusting and folding during the Tertiary. The late Tertiary "extrusion" tectonics of the
Eastern Alps (e.g. Ratschbacher et al., 1991) was mentioned in subchapter 1.2.2.

2.2.2 WESTERN CARPATHIANS


While the western and eastern boundaries of the Western Carpathians are hidden below the Neogene depressions of the Vienna and East Slovak (or Transcarpathian) Basins, respectively, the northern boundary of the Carpathians can be placed between the molasse and the foreland (Figs. 2.5 and 2.10). As the southern border, I adopted an arbitrary boundary with the North Pannonian unit. It is mainly the Hurbanovo-Diósjenő Line, but the Meliata unit outcropping in southernmost Slovakia is also assigned to the North Pannonian system and will be described as such (see Fig. 2.19).

2.2.2.1 STRATIGRAPHY OF THE WESTERN CARPATHIANS

The N-vergent Western Carpathians are traditionally subdivided into two major units. The Outer Western Carpathians include structures formed during the Meso- and Neoalpine orogeny; the Inner Western Carpathians are dominated by Eoalpine units (Fig. 2.10). These large regions are further subdivided into subunits described below from N to S, i.e. beginning with the lower structural unit and ending with the highest thrust sheets.

Similar to the Eastern Alps, the northernmost element is the European foreland with its Mesozoic cover and overlying Miocene molasse (Fig. 2.11). To the S, the Outer Carpathians comprise the Neogene Sub-Silesian and Silesian foreland thrust fold belt and the tectonically overlying Magura Flysch Zone emplaced during the Paleogene. The Malm
Fig. 2.10a. The Western Carpathians. Numbers refer to stratigraphic columns of Fig. 2.11.
Fig. 2.11. Alpine stratigraphy of the Western Carpathians. The location of different units is shown in Fig. 2.10a.
to Oligocene of the Silesian unit includes deep-water clastics. The Magura units are the equivalent of the Rhenodanubian flysch zone of the Eastern Alps and are characterized by flysch and deep-water marls from the Early Cretaceous. The Pieniny Klippen Belt with its complicated internal structure forms the sharp northern boundary of the Inner Western Carpathians, and for most authors it has no counterpart in the Alps. The name refers to the characteristic Jurassic-Cretaceous carbonate klippen piercing Cretaceous-Paleogene flysch. The Klippen Belt is subdivided into several units based on facies differences of generally deep-water Jurassic to Cretaceous strata (e.g. Birkenmajer, 1986, 1988).

The Central Carpathian Paleogene (Slovakia) or Podhale (Poland) flysch basin is restricted to the northern part of the Inner West Carpathians, always to the S of the Klippen belt. This Upper Paleocene to Lower Oligocene succession filled a once continuous basin, but now it is preserved only in erosional fragments (Fig. 2.10). In sharp contrast to the Magura units, the Podhale flysch has no nappes and is only slightly folded (e.g. Nemcok and Neese, 1993). It is comparable to the Szolnok-Transcarpathian flysch belt of the South Pannonian unit (Báldi-Beke and Nagymarosy, 1992; Nagymarosy and Báldi-Beke, in press; cf. Fig. 2.4).

The Eoalpine structural elements of the Inner Western Carpathians are traditionally subdivided into basement and cover nappes. The first group involves a Hercynian crystalline basement with its late Paleozoic to Mesozoic cover (from bottom to top: Tatric, Veporic and Gemenic) nappes. The cover nappes consist of only Mesozoic sequences (Fatricum, Hronic, Silice, Meliata).

In the northern Inner Western Carpathians, the lowermost Tatric nappes are overlain locally by the Fatric cover nappes. The Tatric nappe system with its Triassic to Lower Turonian cover also outcrops in the core of several mountain ranges to the S (Little Carpathians, Inovec, Tribec, see Fig. 2.10). The Fatric nappes are characterized by uniform Upper Triassic continental sediments (Carpathian Keuper). Further subdivision is
made on the character of the Jurassic (Vysoká and Krízna nappes). The Veporic nappes to the S overthrust the Tatric along the Čertovica Line. The reduced Mesozoic cover and the crystalline basement were affected by Eoalpine regional metamorphism. Cretaceous granitoids are also found. At its southern boundary, the Veporic is overthrust by Paleozoic Gemenic units along the Lubeník-Margečany Line. The relationship of the Paleozoic basement of the Gemen region to the overlying Hronic cover nappes (e.g. Choc, Silice) is not clear.

The Mesozoic facies of the Hronic and Silice cover nappes are analogous to that of the Tirolic and Juvavic nappes of the Northern Calcareous Alps and therefore they are regarded as Upper Austroalpine tectonic elements. To the S they also overlie the Meliata unit, which will be described as part of the Northern Pannonian realm.

2.2.2.2 TECTONIC EVOLUTION OF THE WESTERN CARPATHIANS

Similar to the Eastern Alps, the southern edge of the Inner Western Carpathians experienced crustal thinning during the Middle Triassic. This extension progressed to the opening of the Meliata ocean (see page 80). By the end of the Triassic, stable shelf conditions were generally reestablished in areas to the N. During the Jurassic deep-water extensional basins opened up within the Tatric and to the N of it. The latter Vahic oceanic furrow is regarded as the equivalent of the Penninic realm of the Eastern Alps (Plasienka, 1990). Extension characterized the Inner Western Carpathians until the Barremian - Early Albian when basanitic magmatism of mantle origin occurred in the Tatric and Fatric units.

Eoalpine compression began in the Early Albian when flysch was deposited in the Zliechov Trough of the Fatric unit, suggesting the overthrusting of the Veporic to the N. Advancement of the compressional front to the N is shown by the termination of flysch sedimentation in different units. This happened in the Cenomanian of the Krízna unit, in the Cenomanian-Turonian of the southern Tatric and in the Turonian-Coniacian of the
northern Tatric unit. In the Krizna zone, the cover elements were detached from their basement along Werfenian and Keuper decollement levels and accumulated in a collisional accretionary prism. During the Turonian, this pile of cover nappes gravitationally spread over the Tatric to the N (Plasienka, 1990).

The rootless nappe units of the Pieniny Klippen Belt suggest several deep-water Mesozoic basins whose substratum was subducted during Late Cretaceous to Paleogene time. Some of these basins might have originally been deposited on an oceanic substratum (Birkenmajer, 1988). The present-day extremely deformed Klippen Belt was formed during post-Paleogene phases of folding and strike-slip faulting (Birkenmajer, 1986). On the surface, this narrow belt has near-vertical tectonic boundaries everywhere. The N-vergent Magura nappes of the Outer Western Carpathians were formed between Early Oligocene and Middle Miocene. Interestingly enough, during the Middle Miocene the Magura group backthrusted along the Klippen Belt onto the Inner Western Carpathians. Similar S-vergent thrusts are also observed farther to the S (e.g. Biely, 1989).

2.2.3 EASTERN CARPATHIANS


The poorly defined northwestern boundary of the internal Eastern Carpathians lies beneath the East Slovakian (or Transcarpathian) Basin. In the outer flysch belt an arbitrary boundary is drawn to the Western Carpathians at the Polish/Ukrainian border (Figs. 2.5 and 2.11). To the W, the Transylvanian Basin covers the contact with the Tisza block; and to the E, the European foreland flanks the Eastern Carpathians. The southern border with the Southern Carpathians is placed at the Carpathian bend.
2.2.3.1 STRATIGRAPHY OF THE EASTERN CARPATHIANS

Similar to the Western Carpathians, the Eastern and Southern Carpathians (or Dacicides) are traditionally subdivided into two major units. The Outer Eastern Carpathians (Outer Dacides and Moldavides) include structures formed during the Tertiary Meso- and Neoalpine orogeny. The Inner Dacides are dominated by Eoalpine units structurally correlatable with the nappes of the Inner Western Carpathians (Fig. 2.12). These large regions are further subdivided into subunits described below from E to W (structurally from bottom to top).

The Miocene to Recent molasse basin is unusually thick (up to 10 km) on top of the European foreland (Fig. 2.13). The E-vergent Moldavian foreland thrust-fold belt was formed during the Miocene. The internal nappes of the Moldavides (Tarcau, Audia) consist of Lower Cretaceous to Oligocene or Lower Miocene sediments, mostly in flysch facies. The inner nappes of the flysch belt (Convolute, Macla, Ceahlau and Baraolt) also include Tithonian to Lower Cretaceous turbidites.

The Inner Dacides are built up by four major nappe complexes, from bottom to top the Infrabucovinian, the Subbucovinian, the Bucovinian and the Transylvanian units. All are crystalline basement nappes with a Mesozoic cover, but the uppermost Transylvanian nappe is thought to involve an oceanic basement.

The lowermost Infrabucovinian nappes are exposed in small tectonic windows (Fig. 2.12). The whole Triassic is frequently missing from the Mesozoic cover, and most typically Lower Jurassic sediments directly overlie the crystalline basement. The Hettangian-Sinemurian strata are developed in Gresten facies (sandstones with coal seams). The rest of the very reduced Mesozoic is characterized by several stratigraphic gaps attributed to erosion. The same holds for the Subbucovinian nappes outcropping mainly in the northern part of the Inner Eastern Carpathians. The Bucovinian nappes are more widespread and overlie all the above described units in the area between the Rarau
Fig. 2.12a. The Eastern Carpathians. Numbers refer to stratigraphic columns of Fig. 2.13.
Fig. 2.12b. Index map of the Eastern Carpathians.
Fig. 2.13. Alpine stratigraphy of the Eastern Carpathians. The location of different units is shown in Fig. 2.12a.
and Highimas Mts. (Fig. 2.12). In this unit, the most important successions are a Lower Cretaceous (Tithonian - Hauterivian) flysch and above a regional unconformity a Middle Cretaceous (Upper Barremian - Lower Albian) wildflysch. On top of the wildflysch, Upper Albian to Lower Turonian clastics were deposited as a post-tectonic cover.

The stratigraphy of the Transylvanian unit is in part reconstructed from olistoliths found in the above described Bucovinian wildflysch. Three facies units are distinguished (Olt, Highimas and Persani) corresponding to an island arc, ocean and shelf margin, respectively. The Triassic to Jurassic of the Persani unit shows a typical Austroalpine development, with Hallstatt limestones, etc. The island arc sequence of the Highimas unit is characterized by Jurassic basalts.

All of the Inner Eastern Carpathian nappes are covered by an Upper Turonian - Maastrichtian Gosau type (deep-water flysch-like marls) post-tectonic cover. The Paleogene-Neogene basin fill of the Transylvanian Basin also overlies most of these nappes to the W. The Paleogene stratigraphy of the area is almost identical to that of the Hungarian Paleogene basin. The overlying Neogene succession, however, differs from the Pannonian Basin s.s. and includes evaporites in the Middle Miocene (Badenian) strata. This peculiar formation was formed during the Paratethyan salinity crisis (e.g. Steininger et al., 1988).

2.2.3.2 TECTONIC EVOLUTION OF THE EASTERN CARPATHIANS

The early Alpine development of the Inner Eastern Carpathians includes a Triassic crustal thinning period which progresses to the opening of an oceanic basin of the Transylvanian nappe complex (Olt facies zone). More external (i.e. in present-day coordinates to the E) facies zones found in the Infra-, Sub- and Bucovinian nappes display an evolution typical for the European margin, characterized by the absent Upper Triassic and the Gresten facies Lower Jurassic. During the Middle and Late Jurassic one of the
branches of the oceanic Tethys existed in the Transylvanian complex (Highimas zone).

E-directed Eoalpine overthrusting began in the Aptian and lasted until the Late Albian; it involved first the lowermost (outermost) Infrabucovinian and at the end the uppermost (innermost) Bucovinian nappes (Sandulescu, 1988). Since the Bucovinian nappe complex at the top tends to overlie all the other units and spreads to the E to the boundary of the Outer Eastern Carpathians, its gravitational origin is assumed by many (cf. same explanation for the Western Carpathian Fratric). The inner part of the flysch belt was formed during the Late Senonian, but some folding and/or imbrication began as early as the Middle Cretaceous. The Moldavidian overthrusting occurred in the Early and Middle Miocene during three periods, i.e. Burdigalian, Badenian and Sarmatian.

The block-faulted or folded (?) Paleogene fill of the Transylvanian Basin is hidden beneath the Neogene succession. There is a lack of deep seismic data but so far no indication of Middle Miocene extension has been reported. Instead, the Transylvanian Basin looks like a sag basin. Thus the origin of the Transylvanian Basin is not yet adequately understood (cf. Royden, 1988).

2.2.4 SOUTHERN CARPATHIANS


The Southern Carpathians are delimited to the N by the Transylvanian Basin and to the S by the Moesian platform (Fig. 2.14). The eastern boundary with the Southern Carpathians is arbitrary and is commonly placed in the area of the Carpathian bend. To the W and SW there is also an arbitrary geographic boundary to the Balkans drawn along the Danube.
2.2.4.1 STRATIGRAPHY OF THE SOUTHERN CARPATHIANS

The subdivision of the Eastern Carpathians is also applicable to the S-vergent Southern Carpathians (Fig. 2.15). The Outer Southern Carpathians (Outer Dacides) include structures formed during the Tertiary. The innermost Severin nappe is the equivalent of the Outer Dacides flysch belt. The Danubian unit is a crystalline basement nappe between the Severin flysch nappe and the Moldavides. The Moldavides are represented by the Mokranje flysch and the Getic depression. The Eoalpine Inner Dacides of the Southern Carpathians include the Supragetic nappe complex on top and the Getic nappes below. These units are correlated with the Bucovinian and the Subbucovinian - Infrabucovinian nappes of the Eastern Carpathians, respectively. These major units are further subdivided into subunits, described below from E to W (structurally from bottom to top).

The Moesian platform of the European foreland is hidden beneath the Neogene to Recent molasse basin. The molasse also covers the deep Getic depression where a thick (>3 km) Middle Eocene - Lower Miocene sequence accumulated. The Senonian Mokranje flysch is correlatable with the internal flysch belt of the Eastern Carpathian Moldavides.

The Danubian units are subdivided into upper and lower units. The lowermost Schela nappe has an anchimetamorphosed Liassic cover with metasandstones and metapelites and anthracite suggesting a Gresten facies coal-bearing sequence. The Jurassic coals are hard to distinguish from the underlying Carboniferous coals. The Mesozoic of the upper Danubian nappes is also slightly metamorphosed.

The Severin nappe overlies the Danubian unit and outcrops in several narrow stripes, e.g. to the W of the Godeanu Mts. (Fig. 2.14). Most importantly the Severin nappe consists of Upper Jurassic ophiolites (basalts and ultramafics) and the Lower Cretaceous Sinaia flysch.

The Inner Dacides of the Southern Carpathians are further subdivided into an eastern
Fig. 2.14b. Index map of the Southern Carpathians.
Fig. 2.15. Alpine stratigraphy of the Southern Carpathians. The location of different units is shown in Fig. 2.14a.
area (Carpathian bend) that is separated from a western area by the crystalline mass of the Lotru Mts. (Fig. 2.14). The internally imbricated Getic nappe involves mainly a crystalline basement, but in the Carpathian bend and at its western end, its Mesozoic cover is also found. To the E, several nappes are distinguished within the lower Getic (Infrabucovinian) unit (Holbav, Brasov, Leaota, etc.). The common feature of these basement nappes is that the Mesozoic succession begins with Lower Jurassic Gresten facies coal-bearing clastics (Fig. 2.15). Some of the Supragetic nappes (Fagaras, Birsa Fierului) have a reduced Mesozoic cover, but the uppermost Strimba nappe consists entirely of metamorphic basement.

To the W, the Mesozoic cover of the Getic nappes, where present, is also very reduced. The Mesozoic is again characterized by the Gresten facies Lower Jurassic, but in some places the Bathonian - Lower Callovian lies transgressively and directly on top of the crystalline basement. The Upper Jurassic is characterized by pelagic sedimentation, and the Lower Cretaceous developed in a neritic facies. The first post-tectonic cover is an Albian glauconitic sandstone or local bauxites (Hateg basin). The Supragetic nappes in the W are exposed in the Poiana Rusca Mts (Fig. 2.14). These nappes are dominated by a low-grade metamorphosed Paleozoic basement.

2.2.4.2 TECTONIC EVOLUTION OF THE SOUTHERN CARPATHIANS

The very reduced, shallow-water, early Alpine successions of most of the Inner Dacidic units suggest a proximal facies of European affinity. The Jurassic ophiolites of the Severin nappe, however, point to an oceanic basin, overridden during the Cretaceous by internal units.

The Supragetic and Getic nappes involve several Cretaceous periods of deformation. In contrast to the Eastern Carpathian Inner Dacides, where the overthrusting occurred in the Late Aptian - Early Albian, nappe formation in the Inner Dacides of the Southern
Carpathians continued during the Turonian and Late Senonian. The age of the Supragetic overthrust overlying the Getic units in the western area is given in the Rusca Montana Basin (Fig. 2.14) where the Supragetic overthrusts Lower Senonian strata but is covered by Upper Maastrichtian - Paleocene sediments. In the Carpathian bend, the Supragetic units overthrust the Getic nappes along the Holbav thrust and are covered by Upper Maastrichtian sediments. Older overthrusts were frequently reactivated, creating complicated cross-cutting relationships between individual thrust units. Therefore, the overall Eoalpine structure of the Southern Carpathians is much more complex than the structure of the Eastern Carpathians.

The Danubian unit was regarded by some authors as an autochthonous basement of the Inner Dacidian nappe complex. This vast basement unit, however, is also allochthonous and the real autochthonous basement is placed farther to the S of the Getic depression on the Moesian platform. The development of the Late Paleogene Getic depression suggests a foredeep setting.

2.2.5 DINARIDES


The Dinarides are located to the S of the South Pannonian (Tisza) block, and the boundary to the Tisza block is hidden beneath the Neogene basin fill of the southern Pannonian Basin (Fig. 2.16). The easternmost element of the Dinarides is the Vardar zone forming the boundary with the Serbo-Macedonian massif (not shown in Fig. 2.16) and the Southern Carpathians. To the S, the Dinarides are delimited close to the median line of the Adriatic Sea. To the W, the Dinarides gradually merge into the Southern Alps of Italy (see Fig. 2.18) and only an arbitrary boundary can be drawn.
2.2.5.1 STRATIGRAPHY OF THE DINARIDES

The SW-vergent Dinarides are subdivided into four major tectonostratigraphic units. From the SW to NE (from bottom to top), these are the Adriatic, Budva, Dinaric and Internal or Vardar s.l. zones (Fig. 2.16). The lowermost Adriatic zone (or promontory sensu Channel and Horváth, 1976) is built up of a Mesozoic carbonate platform in autochthonous position. It is overridden in the southern Dinarides (S of Dubrovnik, Fig. 2.16) by the Mesozoic deep-water deposits and Paleogene flysch of the Budva zone which represents the northernmost part of the oceanic Pindos-Cukali-Krasta zone of the Hellenides. This unit is overthrust on a large scale (suggested by the Scutari-Spec lateral ramp of Albania, not shown in Fig. 2.16) by the vast Mesozoic platform carbonates of the Dinaric unit. Farther to the NE, the Dinaric nappe complex is overlain by the complicated and poorly known accretionary wedge of the Vardar zone s.l. This comprises the Bosnian unit, the ophiolite belt, the Eastern Bosnian-Durmitor and Drinja-Ivanjica units and the Vardar zone s.s. In the summary stratigraphic chart (Fig. 2.17), the Budva zone is not shown, but the Internal belt is split into the Durmitor and Vardar s.s. zones.

The Adriatic zone outcrops along the Adriatic coast, in the western half of the Istrian Peninsula and on the islands S of Split. It is characterized by an almost uninterrupted carbonate platform succession from the Ladinian to the Middle Eocene. In the Late Eocene, however, flysch was deposited on top of the shallow-water limestones. This flysch is correlatable with the Friuli flysch of the Southern Alps (Fig. 2.17).

The Dinaric unit is subdivided by French authors (e.g. Chorowicz, 1977) into a southwesterly High karst and a northeasterly pre-karst zone. The former spans the area of eastern Istria, the Velebit, Lika and Dinar Mts. (Fig. 2.16), down to Crna Gora (Montenegro). The pre-karst zone is bordered by the Bosnian flysch to the NE. The bulk of the Dinaric unit, similar to the Adriatic unit, consists of a more or less continuous Triassic - Cretaceous platform carbonate sequence. It is interrupted by bauxite horizons
Fig. 2.16a. The Dinarides. Numbers refer to stratigraphic columns of Fig. 2.17.
Fig. 2.17. Alpine stratigraphy of the Southern Alps and the Dinarides. Locations: Figs. 2.16a and 2.18a.
especially in the Cretaceous. In the pre-karst zone above a major unconformity, Upper
Paleocene-Lower Miocene clastics overlie the Mesozoic carbonates.

The Durmitor zone of the Internal belt consists of several fragments of the Upper
Jurassic ophiolitic melange and flysch. This complicated zone overthrusted the Upper
Cretaceous Bosnian flysch. The synthetic stratigraphic column of the Durmitor zone is not
as complete as the columns of the Adriatic and Dinaric units. Tithonian neritic limestones
on top of the succession are unconformably overlain by Berriasian conglomerates grading
into the ophiolitic melange. Locally, Lower Cretaceous intermediate and acidic calc-
alcaline volcanics and granites were found, suggesting island arc magmatism. Similar
volcanics are well documented in the Mures belt of the South Pannonian block.

There are several ophiolite belts in the Internal zone (Ophiolite and Vardar s.s.).
They probably originated from the same oceanic basin to form a single nappe complex but
were separated by subsequent deformation (see Kovács, 1992, for a recent review). The
column of the Vardar zone is reconstructed from the ophiolitic melange of the "diabase-
chert" formation. There is an ongoing debate whether the Vardar Ocean opened as early
as the Middle Triassic or later (cf. Kovács, 1982 and Kozur, 1990). Ladinian pillow lavas
and radiolarites are rare in the Dinarides (Pamic, 1984) but quite frequent farther to the S
in Greece. In any case, the bulk of the Vardar zone s.s. consists of Late Jurassic
serpentinites and gabbros forming isolated blocks within the melange. To the N, the
Vardar zone is covered by the Neogene - Recent succession of the Pannonian Basin.

2.2.5.2 TECTONIC EVOLUTION OF THE DINARIDES

The Dinaric zone with its internal imbricates represents a thin-skinned thrust-fold
belt detached on Triassic evaporites similar to the Apennines on the opposing side of the
Adriatic platform (e.g. Bally et al., 1988). In the Internal zone (e.g. Durmitor), the
Paleozoic and perhaps the crystalline basement are also involved in the overthrusting. This
interpretation suggests large-scale allochthony of the Dinarides, and the palinspastic eastern edge of the Adriatic promontory of Chanell and Horváth (1976) would fall somewhere in the southern Pannonian Basin (in present-day coordinates). In fact, the offshore foreland thrust belt at the Adriatic margin is still active (e.g. Spaic, 1990).

The interpretation of the late, Maastrichtian-Paleogene flysch basins of the internal zones (e.g. Bosnian flysch) is ambiguous. While Pamic (1986) proposed a local subduction zone to explain the flysch sedimentation, Csongos and Vörös (in press) considered the flysch to be deposited in basins in front of major thrust imbricates.

The ophiolitic melange of the Internal zone was formed during the Late Jurassic - Early Cretaceous closure of the Vardar Ocean. A Lower Cretaceous blueschist facies metamorphism also suggests subduction in the Vardar zone s.l. The effect of this metamorphism diminishes to the N and W. The Vardar zone experienced Tertiary NW-trending dextral strike-slip faulting, probably during the Miocene (e.g. Royden et al., 1982). During the Middle Miocene calc-alkaline andesite volcanism occurred in the southern Pannonian Basin and locally Miocene granites were found (Pamic, 1989).

2.2.6 SOUTHERN ALPS

The following brief summary of the geology of the Southern Alps is based largely on the synoptic works of Winterer and Bosellini (1981), Doglioni and Bosellini (1987), Massari (1990), Bernoulli et al. (1990) and the excellent maps of Bigi et al. (1990).

The sharp northern and western boundary of the Southern Alps is the Insubric (Periadriatic) Line including from W to E the Canavese, Tonale, Giudicarie, Pusteria, Gailtal and Karawanka fault segments (Fig. 2.18). To the S, the edge of the Southern Alpine foreland thrust belt is hidden below the Po Plain. To the E, only an arbitrary boundary can be drawn close to the Italian/Slovenian border since the Southern Alpine and Dinaric structures overlap in NE-Italy. The Southern Alps are further subdivided into
a western and an eastern part separated by the Giudicarie fault.

2.2.6.1 STRATIGRAPHY OF THE SOUTHERN ALPS

The present-day structure of the Southern Alps is dominated by a Neoalpine S-vergent thrust belt, but the subdivision into major units is based on early Alpine facies zones (Fig. 2.17). These roughly N-trending zones include from W to E the Lombard Basin, the Trento Plateau, the Belluno Trough and the Friuli Platform (Fig. 2.18).

The stratigraphic sequence of the Friuli Platform is comparable to that of the Adriatic zone of the Dinarides and is dominated by shallow-water carbonates (Fig. 2.17). The Eocene Friuli flysch is unconformably overlain by an Upper Oligocene to Lower Miocene clastic sequence (Massari et al., 1986). Overlying another major unconformity Late Neogene to Recent clastics follow. The Belluno Trough has a distinct deep-water Jurassic facies overlain by Cretaceous pelagic carbonates (Scaglia and Biancone facies).

The Trento Plateau already existed during the early Mesozoic with an uninterrupted Jurassic sequence. This carbonate platform has maintained its relatively elevated and undeformed position. In its southern part (Lessini Mts.) significant Middle Triassic magmatism occurred. This volcanic activity can be seen in all other Southern Alpine zones as numerous tuff intercalations within the Ladinian-Carnian. Overlying an unconformity, the Trento Plateau has an Upper Eocene carbonate cap (Luciani, 1989).

The Lombard Basin shows a distinct deep-water Jurassic development with the deposition of radiolarites during the Bathonian - Oxfordian. The water depth remained considerable throughout the Cretaceous until the Turonian when flysch sedimentation began. This Eoalpine flysch trough had an E-W trend, perpendicular to the direction of the underlying early Alpine zones. Also in the Lombard Basin deep-water clastics (submarine fans) were deposited during the Middle Oligocene - Early Miocene. This peculiar succession is called the Gonfolite Lombarda (e.g. Gunzenhauser, 1985).
Fig. 2.18a. The Southern Alps. Numbers refer to stratigraphic columns of Fig. 2.17.
Fig. 2.18b. Index map of the Southern Alps.
2.2.6.2 TECTONIC EVOLUTION OF THE SOUTHERN ALPS

There is an ongoing debate on the significance of the Middle Triassic magmatism of the Southern Alps (see e.g. Castellarin et al., 1988, for a recent overview), since geochemical studies of the magmatites were not conclusive. The most likely scenario, however, involves aborted rifting (Bechstädt et al., 1978) with a component of left-lateral strike-slip (Doglioni, 1987; cf. Kovács, 1992).

The Jurassic record of sedimentation clearly reflects rifting heralded by the opening of the Penninic Ocean to the W. During the Late Cretaceous this Tethyan passive margin was covered by the Lombard flysch, suggesting a S-vergent Eoalpine thrust-fold belt to the N of the western Southern Alps (Castellarin, 1976, 1977; Doglioni and Bosellini, 1987; Bernoulli and Winkler, 1990). A Mesoalpine period of thrusting is indirectly suggested by the Eocene Ternate Formation (Bernoulli et al., 1988) and by the structural work of Brack (1981) and Schönborn (1992).

After this compressive period the Oligocene "lull" occurred according to Laubscher (1986), who postulated this extensional stage on the base of the occurrence of the Periadriatic intrusions. The radiometric ages of these mainly tonalitic bodies (e.g. Adamello, Bergell, Pohorje) cluster around 30 Ma. Shortly after this extensional (?) period, pronounced backthrusting of the Central Alpine region occurred onto the Southern Alps along the Tonale fault (Heitzmann, 1987).

The Venetian Oligocene basin is regarded as the foredeep of the Dinarides by Massari et al. (1986). NW-trending, SW-vergent Eocene (?) Dinaric thrusts can be traced into the eastern part of the Southern Alps up to the Periadriatic Line (Doglioni, 1987). The Periadriatic Line was initiated during the Late Oligocene (Schmid et al., 1987) and accommodated a significant amount of right-lateral strike-slip during the Early Miocene. The exact amount is subject to controversy (see Chapter 8), but it was at least 160 km.

The present-day S-vergent thrust-fold belt of the Southern Alps began to evolve in
the Middle Miocene. This belt involves only the basement to the north (Laubscher, 1985; Roeder, 1989, 1992). There is an ongoing debate on the amount of N-S shortening across the belt (cf. Doglioni and Bosellini, 1987 and Laubscher, 1990; Schönborn, 1992), but it is certainly considerable (50-100 km). In the eastern part of the Southern Alps, the Neoalpine E-trending thrusts clearly overprint the earlier Dinaric structures.

2.3 PANNONIAN BASIN

The area surrounded by the above described thrust belts is subdivided into two contrasting units based on their pre-Miocene stratigraphy and structure. First, the northern part of the Pannonian Basin (or Pelso) will be discussed since it has an African origin and is correlated with the Southern Alps and the Dinarides. The southern unit (or Tisia) has a European affinity and will be described later.

2.3.1 NORTH PANNONIAN UNIT


The northwestern boundary of the unit is the Rába Line (Fig. 2.19), which represents for many (e.g. Kázmér and Kovács, 1985; Balla et al., 1989; Fülöp, 1990) the ultimate tectonic boundary between the Eastern Alps and the Transdanubian Central Range. (In this introductory chapter this boundary is accepted pro tempore, but in Chapter 5 it will be redefined.) The northern boundary is provided by the Hurbanovo-Diósjenő Line, i.e. the postulated continuation of the Rába Line (e.g. Balla, 1988). From the poorly defined northern termination of the Hurbanovo-Diósjenő Line, the boundary is placed to the N of the Hungarian/Slovakian border to include the Meliata unit. Farther to the E, the pre-
Tertiary basement is poorly known because it underlies the Miocene volcanics of the Tokaj-Zemplén Mts. and the thick Neogene basin fill of the East Slovak Basin. The mostly crystalline Zemplén (or Zemplin) basement is included in the North Pannonian unit. The southeastern boundary is placed in the subsurface at the northern edge of the Szolnok flysch belt (Fig. 2.4). This is the poorly understood Mid-Hungarian Line, running toward the SW to the Hungarian/Croatian border. The Mid-Hungarian Line and the Balaton Line to the N of it (Fig. 2.19) are the postulated eastern branches of the Insubric (Periadiatic) Line. To the W, an arbitrary boundary is placed to the E of the Slovenian Pohorje Mts.

2.3.1.1 STRATIGRAPHY OF THE NORTH PANNONIAN UNIT

Similar to the Southern Alps and Dinarides, a number of early Alpine facies zones are distinguished along a W(WSW)-E(ENE) transect in the North Pannonian unit. These are from W to E the Zala Basin, the Bakony, Vértes - Gerecse, Buda and Bükk Mts. and the Meliata unit (Fig. 2.19).

The stratigraphy of the Zala Basin (Fig. 2.20) is known only from wells since this unit is covered by a thick Neogene basin fill (Körössy, 1988). The deeper part of the Triassic sequence in this unit is not known, but its development is probably similar to the rest of the Transdanubian Central Range (Bakony, Vértes, Gerecse, Buda Mts.) and almost identical to their Southern Alpine counterparts discussed above. The peculiarity of the early Alpine succession in the Zala Basin is the presence of bathyal black shales in the Lower Jurassic comparable to the Lombard zone of the Southern Alps (Kázmér, 1987). Continuous sedimentation was terminated in the Aptian; a Senonian carbonate succession was deposited over a major unconformity, in contrast to the coeval clastic Gosau facies of the Eastern Alps (e.g. Haas, 1985). The Senonian in turn is unconformably overlain by a Middle to Upper Eocene carbonate and clastic sequence including Upper Eocene andesites and tuffs.
Fig. 2.19a. The North Pannonian unit. Numbers refer to stratigraphic columns of Fig. 2.20.
Fig. 2.19b. Index map of the North Pannonian unit.
<table>
<thead>
<tr>
<th>STAGES</th>
<th>ZALA BASIN</th>
<th>BAKONY</th>
<th>VERTES GERECESE</th>
<th>BUDA</th>
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Fig. 2.20. Alpine stratigraphy of the Northern Pannonian unit. The location of different units is shown in Fig. 2.19a.
The Mesozoic and Paleogene of the Bakony unit outcrop in the Keszthely and Bakony Mts. and in the Balaton Highland (Fig. 2.19). In this unit the Permo-Mesozoic succession unconformably overlies Hercynian low-grade metamorphics. The Anisian-Ladinian sequence shows the same volcanoclastic development as in the Southern Alps. The thick (>3 km) Upper Triassic carbonate platform strata is overlain without interruption by deep-water Jurassic radiolarites. By the beginning of the Cretaceous shallow-water carbonates were again deposited. Compared to the neighboring units the Bakony unit was a relatively elevated submarine plateau during early Alpine time, perhaps correlatable with the Trento unit of the Southern Alps (Kázmér, 1987). An unconformity-bounded Aptian and Albian - Cenomanian sequence follows. The Senonian Basin in the Bakony Mts. is identical to that of the Zala Basin. A Late Eocene transgressive sequence is found above another unconformity with erosional top. It is overlain by Upper Oligocene to Lower Miocene continental clastics.

The early Mesozoic succession in the Vértes-Gerecse unit is comparable to the previously described North Pannonian units. However, a significant deviation from this pattern is found in the Early Cretaceous sequence of the Gerecse Mts., where the Tithonian neritic limestones are overlain by sandstones, marls and finally flysch deposited up to the Barremian. These are followed by deep-water Aptian - Lower Albian clastics. This Eoalpine sequence finds its counterpart in the Internal zone of the Dinarides outcropping in the Ivanjica Mts. (see Fig. 2.16).

Due to subsequent erosion the Mesozoic of the Buda Mts. is very reduced. An important difference compared to the other Transdanubian Triassic developments is the very thick (>1200 m) Ladinian dolomite deposited on a carbonate platform. The Paleogene sequence in the southern part of the Buda unit ranges from the Upper Eocene up to the Lower Miocene. This Paleogene is almost identical to that of the Slovenian Paleogene Basin in the northern Dinarides (e.g. Nagymarosy, 1990b; see Fig. 2.4). Over
an unconformity, the sequence continues to the Middle Miocene with only a small hiatus.

While the early Alpine successions of the Transdanubian units show a clear South Alpine affinity, the Bükk and Meliata units show a Dinaric-type development (e.g. Balogh, 1964). The boundary of the Buda unit with the Bükk unit is located just to the NW of the Danube bend, since in the small hill of Csővár (Fig. 2.19) Bükk-type Triassic crops out. The Bükk unit consists of several S-verging nappes (Csontos, 1988). The lowermost one is exposed in the Bükk Mts. (Fig. 2.19), and is called the Bükk paraautochthon. The whole early Alpine sequence is low-grade metamorphosed (Árkai et al., 1985). The Lower Triassic is developed in a neritic carbonate facies with characteristic black dolomites. The Middle Triassic here is also volcanogenic with calc-alkaline intermediate to acidic metavolcanites (Cros and Szabó, 1984). By the end of the Carnian, the carbonate shelf was drowned, and deep-water sedimentation with turbidites and olistostromes persisted during Late Triassic - Jurassic times. There is a thick (>2 km) Upper Eocene - Lower Miocene sequence in northern Hungary unconformably overlain by Middle and Upper Miocene volcanics and clastics of the Pannonian Basin.

The Meliata unit outcropping beneath the Silice unit of the Inner Western Carpathians in southern Slovakia (Fig. 2.19) tectonically overlies the Bükk nappe system. The Meliata unit is an isolated klippe on the southwestern part of the Bükk Mts., the Szarvaskő nappe (Fig. 2.19). This unit is characterized by deep water Triassic-Jurassic strata with ophiolites in the Middle Triassic (Darnó Hill) and in the Upper Jurassic (Szarvaskő). This unit is correlated with the Vardar zone of the Dinarides (e.g. Kovács, 1982, 1984, 1992; Balla, 1987c; Csontos, 1988).

2.3.1.2 TECTONIC EVOLUTION OF THE NORTH PANNONIAN UNIT

The widespread Middle Triassic volcanism in Transdanubia is regarded as the manifestation of the "aborted rifting" of Bechstädt et al. (1978). To the NE, in the Bükk
unit, the geochemical character suggests a convergent margin setting (Kovács, 1992).

The pelagic Jurassic sedimentation in the Transdanubian units suggests an oceanic realm to the N(NW) in present-day coordinates (Galácz et al., 1985). In Northern Hungary, i.e. in the Bükk Mts., the Vardar ocean opened up as early as the Middle Triassic and existed as an oceanic domain until the Jurassic. The Early Cretaceous flysch in the Gerecse Mts. appears related to the closing of the Vardar ocean (Sztanó, 1990). This closing was accompanied by the southerly overthrusting of numerous Eoalpine nappes of the complicated Bükk-Meliata-Silvice-Choc systems. The emplacement of the Meliata - Szarvaskő nappes over the Bükk paraautochthon occurred during the Early Cretaceous (115 Ma K/Ar age, Árva-Sós et al., 1986) associated with LT/HP metamorphism (Árka, 1983). S-vergent nappe movements continued until the Middle Cretaceous (90 Ma K/Ar age, Árva-Sós et al., 1986).

In contrast to the traditionally accepted Eoalpine nappe structure of Northern Hungary (e.g. Balogh, 1964), the effect of the Eoalpine stage in the Transdanubian region is commonly considered negligible (e.g. Balla, 1992b). The Middle Cretaceous (Aptian - Albian) compression caused only local thrusting in the Bakony Mts. and formed a major syncline in the axis of the Transdanubian Central Range. Based on subsurface data, however, Horváth and Rumpler (1984), Horváth et al. (1987) suggested the large-scale allochthony of the Bakony Mts.

The Senonian of the Zala and Bakony units and the whole Paleogene of the Northern Pannonian are considered to be a post-tectonic, epicontinental or molasse-type sequence (e.g. Haas, 1979, 1987 and Balázs et al., 1981; Nagymarosy, 1990a, respectively). The Paleogene basin was also classified by some as a pull-apart feature related to regional dextral strike-slip (e.g. Báldi and Báldi-Beke, 1985; Royden and Báldi, 1988; Fodor et al, 1992; Csontos et al., 1992).
2.3.2 SOUTH PANNONIAN UNIT (TISIA)


The boundaries of the Southern Pannonian realm are almost entirely covered by Neogene sedimentary fill from the southern part of the Pannonian Basin (Fig. 2.21). To the N, this unit is bordered by the enigmatic Mid-Hungarian Line beyond the northern edge of the Szolnok flysch belt. To the E, the boundary follows the southern edge of the Botiza Klippen to the N of the Preluca Mts. To the S, the boundary to the Eastern Carpathians is poorly known and is placed in the N-S axis of the Transylvanian Basin. The contact with the Southern Carpathians is rather sharp and runs between the southern Apuseni Mts. and the Poiana Rusca Mts. To the S, the boundary of the Vardar zone of the Dinarides is ill-defined and is drawn somewhat arbitrarily. To the SW, the boundary is best placed in the NW-trending axis of the Sava Basin.

2.3.2.1 STRATIGRAPHY OF THE SOUTH PANNONIAN UNIT (TISIA)

In contrast to the Northern Pannonian block, the Tisia block is subdivided into roughly E-W trending units related to dominantly N-vergent Eoalpine deformation (e.g. Kovács, 1982). They include from N to S (or from bottom to top) the Mecsek-Szolnok, the Villány - Bihar, the Békés - Lower Codru, the Upper Codru - Biharia and the Mures units (Fig. 2.22) with additional subunits where further subdivision was possible.

The lowermost Mecsek-Szolnok unit outcrops only in the Mecsek Mts. of Southern Hungary (Fig. 2.21). The early Alpine sequence starts with a thick (>2 km) Permian continental succession unconformably overlying a metamorphic Hercynian basement (Fig. 2.21). Above a neritic-lagoonal Lower and Middle Triassic, the Upper Triassic is again characterized by continental sedimentation. This development and the thick Liassic coal-
Fig. 2.21b. Index map of the Southern Pannonian (Tisia) unit.
Fig. 2.22. Alpine stratigraphy of the Southern Pannonian (Tisia) unit. The location of different units is shown in Fig. 2.21a.
bearing sequence in Gresten facies clearly suggest the position of this unit on the European margin of the later Tethys. The Jurassic Ammonoides-fauna also shows the same European affinity (Géczy, 1973). The Middle and Upper Jurassic record the deepening of the sea below the CCD by the Oxfordian when radiolarites were deposited. The Tithonian limestones are overlain by Lower Cretaceous basalts. A Senonian clastic sequence follows above a major unconformity. The most peculiar part of the Mecsek unit is the Paleocene - Oligocene Szolnok flysch (see Szepesházy, 1973, for a summary) found in wells in northeastern Hungary (Fig. 2.21). This belt continues in the subsurface to the E to Romania, where it is called the Maramures or Borsa flysch. In the northern part of the Eastern Carpathians, it outcrops as the Transcarpathian flysch. The Szolnok flysch is not a typical flysch (e.g. Juhász, 1968) although it also has deep-water clastic facies. The Szolnok flysch represents a stratigraphically discontinuous succession (e.g. Báldi-Beke and Nagymarosy, 1992). It is best correlated with the Central Carpathian Paleogene or Podhale flysch basin (Nagymarosy and Báldi-Beke, 1993; see Fig. 2.4).

The Villány-Bihor unit outcrops in the Villány Mts. of S-Hungary and in the Bihor area of the Apuseni Mts. of E-Romania. The physical connection between these regions beneath the Great Hungarian Plain is constrained by numerous wells (e.g. Szepesházy, 1979). The early Alpine development of this zone differs significantly from those of the neighboring units, since it is characterized by an incomplete sequence with condensed strata, suggesting a platform flanked by basins to the N and S. There is a hiatus in its upper part of the continental-to-neritic Triassic sequence. The overlying Gresten-clastics pass upward into an overall deepening, but incomplete Jurassic carbonate sequence. A significant change occurred in the Oxfordian when continuous and thick carbonate deposition started. The Cretaceous is again characterized by discontinuous sedimentation. The distribution of Senonian Gosau-type clastics is seemingly influenced by the thrust front of the Codru nappe system to the S. The peculiar Senonian acidic "banatite"
volcanism of the eastern part of the Villány - Bihor zone is best exposed in the Bihor Mts. The Campanian and Maastrichtian are characterized by tuffs, rhyolitic lava-flows and subvolcanic intrusions. The banatic volcanicism continued into the Paleocene.

The Lower Codru unit outcrops in the Papuk Mts. in Croatia and in the Apuseni Mts. of Romania, where the unit is subdivided into several nappes. These are not yet differentiated in the Papuk Mts. or in the subsurface of the Great Hungarian Plain. The most characteristic feature of this unit is the Lower Jurassic - Lower Cretaceous flysch of the uppermost Urmat nappe.

The Upper Codru décollement nappes consist of Mesozoic sequences only. The Biharia nappes, however, are basement nappes composed of Hercynian metamorphics. The Mesozoic facies of the Upper Codru nappes is similar to that of the Lower Codru nappes but differs by having Middle Triassic pelagic sediments.

The uppermost Mures nappe complex is exposed only in the southern part of the Apuseni Mts. This zone projects below the Transylvanian Basin to the NE. The relationship to the Transylvanian unit of the Inner Dacides is postulated because of the similarity of these units. The Mures belt continues beneath Vojvodina to the WSW where it joins the Vardar zone s.s. of the Dinarides. The Mures complex has a complicated structure involving a large number of nappes. Based on the stratigraphy of these nappes a magmatic arc and a back-arc basin are reconstructed. Upper Jurassic - Lower Cretaceous basalts with clear MORB affinity are found only in the Fenes nappe system. During the Late Cretaceous several unconformity-bounded clastic sequences were deposited. Banatite tuff intercalations are abundant in the Senonian deep-water clastic sequence.

2.3.2.2 TECTONIC EVOLUTION OF THE SOUTH PANNONIAN UNIT (TISIA)

The early Alpine facies of various units indicates in present-day coordinates a continental margin to the N and a deep oceanic basin to the S. The subduction of this
oceanic basin began in the Early Cretaceous.

The N-vergent Eoalpine structure of the Mecsek-Szolnok and Villány - Bihor units differs from the southern Codru units. The former are regarded by many as an imbricated, N-vergent thrust belt without real nappe structures (e.g. Wein, 1969; Némedi-Varga, 1983), whereas the Codru and Mures complexes are organized in a still N-vergent complicated nappe pile (e.g. Bleahu, 1976). The thin-skinned Upper Codru nappes were detached from the Biharia crystalline basement. The emplacement of these cover nappes onto the Lower Codru nappes, according to most authors, took place by gravity gliding.

The topmost Mures belt displays a bilateral vergency with S-vergent nappe units on its southern flank. The Mures unit is interpreted as a Late Jurassic - Early Cretaceous oceanic island arc and a corresponding back-arc basin formed to the S of the Apuseni Mts. During the Neocomian, these elements were incorporated into an accretionary wedge. This wedge was further deformed during the Albian and Senonian. The Upper Cretaceous - Paleocene banatites postdate the nappe movements since they crosscut all the thrust contacts.

The structure of the Mures belt also records a Tertiary period of strike-slip faulting. The en-échelon arrangement of crystalline domes and intervening synclines of cover sequences suggests right-lateral shear sense. The thickening of Middle Miocene sediments towards the axis of the synclines dates the fault activity.

The present-day structure of the South Pannonian unit, similar to the North Pannonian unit, is dominated by Neogene extensional features (see Royden and Horváth, 1988, for a review). This latest stage in the structural evolution of the South and North Pannonian units is best illustrated by the regional reflection seismic sections of the next chapter.
CHAPTER 3
REGIONAL SEISMIC REFLECTION SECTIONS ACROSS THE HUNGARIAN PART OF THE PANNONIAN BASIN

Following the brief introduction in Chapter 2 of the Alpine geology of the Pannonian Basin and the surrounding mountain belts, this chapter shows reflection seismic data from the central part of the Pannonian Basin. This area of Hungary is illustrated by 18 regional seismic cross sections (Fig. 3.1). The first six are located in Western Hungary trending generally NW-NNW. As discussed in Chapter 2, these lines may be regarded mainly as dip lines with respect to Eoalpine and Neoalpine structures. In Western Hungary, the unevenly spaced seismic coverage did not permit a set of strike lines to be constructed connecting the dip lines. In Eastern Hungary, however, the very dense seismic grid allowed the construction of a perpendicular, NE-ENE trending set of regional seismic lines (Fig. 3.1). The 12 regional lines cover the whole area of the Great Hungarian Plain, crossing all the major tectonic units and "lines" of Hungary described in Chapter 2 (Fig. 3.2a,b).

The individual cross-sections located on Fig. 3.1 were constructed by splicing a number of shorter seismic reflection profiles, listed in Appendix D. These sections are of different vintages, representing different data acquisitions, processing and displays. The data, however, come mostly from the oil industry, principally from the Geophysical Exploration Company of Hungary. Supplemental data were obtained from an academic organization, the Eötvös Lóránd Geophysical Institute.

To give a unified interpretation of these different data sets, I chose to present line drawing interpretations of the seismic data. Since the regional lines are very long (up to 300 km) I applied a two times vertical exaggeration to the standard time scale of oil industry lines. The academic lines were adjusted accordingly. The interpretation was also aided by the projection of more than 200 wells (listed in Appendix D) into the sections.
Fig. 3.1. Index map of regional seismic lines and supplementary seismic sections in Hungary.
Fig. 3.2a. Depth of pre-Tertiary basement in Hungary with the grid of regional seismic lines shown in Fig. 3.1.
Fig. 3.2b. Pre-Tertiary subcrop map of Hungary with the grid of regional seismic lines shown in Fig. 3.1.
Fig. 3.3 shows the legend for the regional seismic lines. Only the most important horizons were correlated at this scale. They were subdivided into three major units: pre-rift, syn-rift and post-rift, with respect to the Neogene extension of the Pannonian Basin. A further subdivision was made within these successions (Fig. 3.3).

At this point it is unavoidable to introduce some regional stage names that are commonly used in the Tertiary of the so-called "Paratethys" of Central and Eastern Europe (for a detailed discussion see Appendix B). The realm of the Paratethys includes the major Late Tertiary basins to the N of the Alps and the Dinarides, extending from the Swiss Molasse basin in the W, all the way to the Caspian Sea in the E (e.g. Rögl and Steininger, 1985; Steininger et al., 1988). Since the end of the Eocene the Paratethys showed repeated periods of isolation from the world sea, due to uplift of certain Alpine mountain ranges. This separation of the Paratethyan basins is mainly reflected in the increasing endemism of the aquatic biota. The first appearance of endemic biota occurred in the beginning of the Oligocene (Eoparatethys), followed by an Early Miocene (Mesoparatethys) and Late Miocene (Neoparatethys) pulse of endemism. In between these periods, marine connections with the Mediterranean region were reestablished, but the Paratethys realm remained a distinct paleogeographic province (see e.g. Baldi, 1989 for a review). Periodic endemism has prevented biostratigraphic correlation of the Paratethyan sedimentary successions with the standard stages of the Mediterranean. Therefore regional stages were introduced, also within the Central Paratethys, i.e. the Carpatho-Pannonian region. Almost all of the Paratethys regional stages are defined by transgressive-regressive facies cycles sensu Vail et al. (1992) with specific faunal assemblages. The description of the Paratethyan sedimentary succession (Oligocene to Recent) in the Pannonian Basin in terms of sequence stratigraphy can be found in Appendix B. Based on this sequence stratigraphic approach, a correlation scheme between the standard Mediterranean and regional Central Paratethyan stages is shown in Fig. 3.4.
Fig. 3.3 Legend of regional seismic sections.
Fig. 3.4. Correlation of the regional Central Paratethys stages with the standard Mediterranean stages, see Appendix B for more details.
On the simplified line drawing interpretations of this chapter not all the reflectors are shown in order to present a legible display. In addition to the line drawings of the regional lines, many seismic sections are also shown for specific illustration (Fig. 3.1). The compilation of 35 seismic sections in this chapter represents a "seismic mini atlas" of published reflection seismic data in Hungary.

The regional lines are shown in three separate panels (Panels 1, 2, 3), named by upper case letters (A-R) going from W to E and from N to S. Pronounced changes (>45°) in the direction of the section are indicated. The supplementary seismic lines are either parts of a given regional section or sections near it. These lines are named with the combination of the upper case letter of the corresponding regional section and an additional lower case letter (Fig. 3.1). Please note that the scale of the supplementary seismic illustrations is always different from that of the regional lines.

3.1 WESTERN HUNGARY, DIP LINES

These regional lines (Panel 1) cross the area of the Zala, Drava, and Danube Basins and the Transdanubian Central Range. The gaps in the sections on lines B, C, D and F are due to lack of seismic data over Lake Balaton (Fig. 3.1). In addition, some regions (e.g. Bakony Mts.) did not permit detailed interpretation due to poor data quality.

In this area, the Pannonian part of the post-rift succession was not subdivided into subsequences since progradational units are not well developed. In the pre-rift unit, however, a thick Senonian basin can be delineated as well as an overlying Paleogene sequence.

3.1.1 REGIONAL SECTION A

The northwestern end of the regional section between 0 and 22 km is based on the Aa line (Fig. 3.5) published by Rumpler and Horváth (1988). The basement high to the
Fig. 3.5. A seismic line of Rumpler and Horváth (1988): stacked (upper), migrated section (middle) and line drawing interpretation (lower). For location see Fig. 3.1.
NW is the South Burgenland swell and is made up of low-grade metamorphosed Paleozoic rocks. To the SE, however, the basement consists of unmetamorphosed Mesozoic rocks of the Transdanubian Central Range. The fault in between is a low-angle normal fault corresponding to the enigmatic Rába fault (see 4.10.1 subchapter for a detailed discussion) of Rumpler and Horváth (1988). A significant amount of sinistral strike-slip was postulated by these authors along its oblique-slip fault plane.

The seismic facies right on top of the fault plane (Fig. 3.5) suggests syn-rift coarse clastic aprons of Middle Miocene (Karpatian and Badenian) age, which were indeed drilled along strike (e.g. Kotormány). The very pronounced syn-rift/post-rift boundary in the basin center at 15 km coincides well with the Badenian/Sarmatian boundary correlating the Miocene horizons from well #1 at 21 km. The post-rift Sarmatian and the lower part of the Pannonian sequence clearly onlap the basement high. Even the youngest reflectors are truncated near the top of the section, pointing to the young uplift of the NW side of the section relative to the basin.

To the SSE the regional section enters the deeper part of the Zala Basin, at Zalalővő (well #2). Faint S-dipping reflectors in the pre-Miocene basement between 40 and 45 km indicate its complex internal deformation which is confirmed by nearby wells (Ze-2, Ze-3 and Brn-1) in which repeated sections of Jurassic/Triassic rocks were found (see Chapter 5). Well #5 encountered Upper Eocene andesite and tuff in the Zalatárnomok-Bak Trough. In this E-W trending trough also a thick (up to 1000 m) Senonian sequence is preserved (e.g. Dubay, 1962; Körössy, 1988). To the S, wells #7 and #8 encountered steeply dipping (~50°) Triassic carbonates in the basement. The so-called Balaton Line (e.g. Dubay, 1962) can be projected into regional section A at about 70 km.

Wells #9-#12 were drilled on the Budafa anticline trending perpendicular to the section at 75 km. The Ab profile shown in Fig. 3.6 is located just some kms to the E (Fig. 3.1). This line published by Dank (1988) shows the very young age of the anticline.
This anticline and several others in SW-Hungary and in the adjacent areas of Slovenia and Croatia (e.g. Dank, 1962) represent the so-called Sava folds that are traditionally regarded as Early Miocene compressional features (cf. Savian phase of Stille, 1953). A closer look at the thickness relations of the syn-rift strata using abundant borehole data (e.g. Bodzay, 1966) reveals the inversionsal origin of the structure. In fact, a Middle Miocene half-graben with a S-dipping master fault on its northern margin was inverted during Quaternary-Recent times. The internal details of the Budafa anticline could not be resolved on the regional section.

The regional section enters the area of the Letenye subbasin S of 85 km. Here well #14 encountered Eocene andesite (Dank and Fülöp, 1990) below Karpatian sediments. Farther to the S, in the area of wells #15, #16 and #17, the Pannonian succession is again clearly folded above the poorly known pre-Tertiary basement. At the southern end of regional section A, well #18 in the Dráva Basin found a more than 2400 m thick syn-rift (Middle Miocene) succession. The same well also reached the pre-Tertiary basement consisting of Senonian marl and limestone. This area, however, is more typically characterized by a polymetamorphic basement (see Fig. 3.2b).

3.1.2 REGIONAL SECTION B

While the NW end of regional section A is located in the transitional area between the Zala and Danube Basins, section B starts in the latter. The northwestern end of regional section B between 0 and 23 km is based on the Ba line (Fig. 3.7) published by Tari et al. (1992b). On the northwest side of the section, the South Burgenland basement high is made up of epimetamorphic greenschists, according to borehole evidence (well #1). These Jurassic-Early Cretaceous rocks are the subsurface continuation of rocks outcropping at the Austrian-Hungarian border in the Penninic window of Rechnitz (e.g. Pahr, 1980). In contrast, on the southeast side of the section, anchimetamorphic Paleozoic
Fig. 3.7. The Ba seismic reflection line published by Tari et al. (1992). Migrated section (a) and its interpretation (b). For location see Fig. 3.1. The basement consists of epimetamorphic greenschists on the northwestern side of the section, representing the subsurface continuation of Jurassic-Early Cretaceous rocks outcropping in the Penninic window of Rechnitz. The well on the southeastern side of the profile bottomed in anchimetamorphic Paleozoic rocks (Graz Paleozoic). The low-angle tectonic contact between these tectonic units corresponds to a Cretaceous major overthrust plane. During the middle Miocene the same fault plane reactivated as an extensional detachment fault, along which the metamorphic core complex of Rechnitz (Rohonc) was uplifted and the asymmetric trough of middle Miocene through Pliocene in age subsided.
rocks (Árkai and Balogh, 1989) were drilled at 24 km (well #2). The low-angle tectonic contact between these two different types of basement unit corresponds to a two-phase dip-slip displacement. Cretaceous (Eoalpine) overthrust movements are responsible for the older-on-younger relationship, while the sharp change in metamorphic grade is seen to be the result of Miocene extensional detachment faulting, reactivating a pre-existing thrust plane (Tari and Bally, 1990). The large normal offset along this low-angle normal fault (see Chapter 5) can be dated by clastic fan deposits which accumulated due to the tectonic denudation of the footwall block (cf. Fig. 3.5). The age of these coarse clastics is Middle Miocene (Karpätian).

From about 35 km on, the pre-Tertiary basement is made up of the unmetamorphosed Mesozoic rocks of the Bakony Mts., according to boreholes #3-9. A package of reflectors within the basement dips gently below the Bakony Mts. These reflectors are mappable in this area and regarded as fault plane reflections of an Eoalpine thrust (see Chapter 5). After depth conversion this thrust plane has a more pronounced dip to the SE.

The gap between 65 and 101 km is due to a lack of seismic reflection data in the area just to the W of the Keszthely Mts (Fig. 3.1). Fig. 3.8 shows, however, the Bb line which is located some 20 kms to the W of this gap (Fig. 3.1). This section from Pogácsás et al. (in press) illustrates the internal structure of the Nagylengyel oil field. The Senonian and overlying volcanoclastic Eocene succession was blockfaulted and slightly compressed. The syn-rift succession is very thin in this area.

From 101 km onwards, the Neogene basin deepens to the S. The basement is made up of medium-grade micaschists in borehole #10. Farther to the S, Körössy (1988) reported a Triassic basement in wells #11 and #12; however, Árkai (1987) described the rocks in well #11 as micaschist and gneiss. Thus the latter author placed the Balaton Line between wells #11 and #12, at 115 km of regional section B.
Between 110 and 145 km, regional line B coincides with line Bc shown with an opposite orientation in Fig. 3.9 published by Rumpler and Horváth (1988). The deeper part of the graben between 115 and 130 km is filled by Lower Miocene volcanites based on borehole information (see regional section C below), magnetic anomalies (e.g. Posgay, 1966) and seismic character. The core of the young Vése anticline between 130 and 140 km shows some reflectors, and wells #13 and #14 drilled thick syn-rift Miocene strata. While there is again no doubt about the very young age of the anticline, Rumpler and Horváth (1988) suggested that the main folding occurred during the Middle Miocene. The internal geometry of the structure, however, suggests that a syn-rift half-graben was inverted subsequently. Small, but deep half-grabens can be found indeed in the nearby Kadarkút area (see regional section C).

The poorly defined Mid-Hungarian Line (Fig. 3.2b) may be identified as the master fault at 140 km, since southward the basement is made up of Mecsek-type crystalline rocks based on the nearby Kutás wells (Körössy, 1989). Farther to the S the pre-Tertiary basement is not known due to Middle Miocene volcanics (wells #15 and #16). However, wells #17, #18 and #19 again encountered the high-grade polymetamorphic basement of the Mecsek unit. The structural relation to the basement of the Görgeteg-Báböcsa anticline beneath wells #18 and #19 is not clear, but its very recent formation is again obvious. At the end of the section an Upper Carboniferous very low-grade sequence was found in well #20.

3.1.3 REGIONAL SECTION C

The northwestern end of regional section C between 0 and 50 km coincides with the Ca line published by Hobot et al. (1990). This line is reproduced and reinterpreted in Plate 1. The pre-Tertiary basement is made up of very low- to low-grade metamorphosed Paleozoic rocks up to 37 km according to the wells in three basement
Fig. 3.9. Bc line of Rumpler and Horváth (1988), with stacked, migrated and line drawing versions. For location see Fig. 3.1.
highs (Bük, Ölbö and Ikervár, respectively). Just 2 km to the SE of well #2, however, borehole #7 penetrated unmetamorphosed Senonian and Triassic rocks belonging to the Mesozoic of the Bakony Mts. The enigmatic strike-slip fault of the Rába line (see 5.10 subchapter for a detailed discussion) is traditionally placed between these two wells separating contrasting basement units (e.g. Körössy, 1987). The seismic data (Plate 1), however, do not reveal any offsets which might be attributed to a subvertical fault. In fact, the projection of the Rába fault as it is redefined in this thesis (see Chapter 5) reaches the top of the basement at 30 km as a bundle of SE-dipping reflectors.

The undulating reflectors in the basement between 5 and 10 km were interpreted by Hobot et al. (1990) as related to a SE-vergent thrust. Rotated fault blocks, however, bounded by low-angle normal faults shown in regional section C offer a more viable alternative. These low-angle faults appear detached on a gently SE-dipping surface marked by a set of strong reflectors (Plate 1). This surface is interpreted to be the top of Penninic metamorphites and thus identical to that shown in regional section B between 0 and 25 km. Well #1 penetrated Middle Miocene coarse clastics between the fault blocks dating the extensional detachment faulting.

The continuation of regional section C to the SE coincides with the Cb line (Plate 2) between 50 and 115 km down to Lake Balaton (Fig. 3.1). The Neogene basin becomes gradually thinner between 50 and 95 km and the pre-Tertiary basement outcrops in the Bakony Mts between 95 and 115 km. The Pannonian reflectors dipping to the NW indicate the young uplift of the Bakony Mts.

In the area of the outcropping Mesozoic, the data quality is rather poor; however, a number of prominent reflectors can be traced in the basement suggesting NW-directed thrust planes and a décollement level at about 2 s TWT time depth 55 and 95 km.

The gap between 115 and 128 km is due to Lake Balaton (Fig. 3.1). Several boreholes (#8, #9 and #10) show the location of the Táska-Buzsák basement high between
135 and 140 km. The Balaton Line is placed just to the S of it, since Dinaric type Late Paleozoic - Triassic (?) limestone and marl were reported from well #11 by Árkai (1987). This enigmatic line is defined by an E-W trending zone of structural disturbance which is revealed by many of the wells drilled in the area (e.g. Sztrákos, 1975). For example in the plane of regional section C wells #8 and #11 penetrated repeated sections of Triassic carbonates and Lower to Middle Miocene sediments (e.g. Balla et al., 1987; Körössy, 1990). The available seismic data, however, do not reveal any of these structures.

The chaotic reflectors in the syn-rift fill of the Mezőcsokonya Trough represent Lower and Middle Miocene intermediate and felsic volcanites according to wells #12 and #13. In well #14 Miocene rhyolites (ignimbrites?) were described by Körössy (1990). The Kaposföi basement high between 165 and 172 km is made up of Mecsek-type polymetamorphics based on nearby wells. Thus the Mid-Hungarian line should be expected to the N of this basement high at about 160 km(?), in this section obscured by thick volcanic cover.

The Cc line shown in Fig. 3.10 is trending parallel with regional section C between 163 and 185 km, but it is located about 8 km to the W. Rumpler and Horváth (1988) interpreted this section in terms of extensional half-grabens in the Kadarkút area. The main trough is filled with a thick Middle Miocene syn-rift succession (>1300 m in well #15) and bounded by a N-dipping master fault. Interestingly enough this deep graben was not inverted (cf. Vése anticline of regional section B!).

The nature of extension is much less clear to the S in the deep Dráva Basin from 195 km up to the end of regional section C. The very young folding so characteristic of the sections described above can also be observed here in the Neogene sequence, although with less intensity. The pre-Tertiary basement is made up of polymetamorphics and very low-grade upper Paleozoic clastics.
Fig. 3.10. Cc seismic line published by Rumpler and Horváth (1988): stacked (upper), migrated (middle) and line drawing interpretation (lower). For location see Fig. 3.1.
3.1.4 REGIONAL SECTION D

The northwestern end of regional section D between 0 and 58 km coincides with the Da line (Fig. 3.11) published by Posgay et al. (1986). This crustal reflection section can also be found in Plate 3 with an alternative interpretation. The pre-Tertiary basement is made up of Paleozoic crystalline rocks between about 0 and 28 km based on the nearby outcrops of the Sopron Mts. (Fig. 3.1) and borehole data close to the section (Pinnye and Csapod; Körössy, 1987). The Pinnye basement high at 18 km is flanked by two half-grabens. The one to the NW is called the Nagycenk Basin, and the one to the SE is the Csapod Basin (Ádám et al., 1984). The basement culmination at 40 km is the Mihályi high and is made up of low-grade metamorphosed Paleozoic rocks (Balázs, 1971; 1975).

Both the Pinnye and Mihályi highs are bounded by major low-angle normal faults on their SE side. In fact, the fault which bounds the Mihályi high is the Rába fault of a number of authors (see Chapter 4) and it was regarded by some as a subvertical oblique-slip fault (e.g. Horváth et al., 1987). Their interpretation is best illustrated by the Db line shown in Fig. 3.12. This seismic example is located roughly parallel to and SW of regional line C between 35 and 80 km. Rumpler and Horváth (1988) proposed the steepening of this fault plane with increasing depth.

In any case, to the S of this fault the pre-Tertiary basement is covered by an Upper Cretaceous basin succession. The thickening of Senonian strata towards the Rába fault (Fig. 3.12) is not as evident on regional section D. There, instead, the Middle Miocene syn-rift sequence shows a pronounced thickening in the Kenyeri subbasin of the Danube Basin at about 50 km.

The pronounced reflector packages within the pre-Senonian basement suggest a number of NW vergent thrust faults emanating from a décollement level. The décollement has a seemingly gentle dip to the NW; however, this dip is more pronounced after correction for the velocity pull-up. Moreover, a simple restoration of the pre-rift geometry
Fig. 3.11. Detail of the Da deep seismic reflection profile from Posgay et al. (1986). For location see Fig. 3.1.
Fig. 3.12. Db line of Rumpler and Horváth (1988), with stacked, migrated and line drawing versions. For location see Fig. 3.1.
also increases its dip to the SE between about 60 and 100 km.

The package of strong basement reflectors interpreted as a detachment level has been known for about a decade. In addition, this level traced on several seismic sections in the area appears to correlate with a conductivity anomaly in the basement (e.g. Takács, 1968; Ádám et al., 1984; Horváth et al., 1987). Fig. 3.13 shows the Dc line (Albu et al., 1983) located to the NW of regional line C (Fig. 3.1). According to the magnetotelluric sounding in the center of Dc line, the Senonian strata have a 38 ohm/m resistivity, in contrast to the underlying mainly Triassic carbonate succession with a 300 ohm/m. The anomalously high conductivity zone (1 ohm/m!) is located at 5.8 km depth and coincides with a band of prominent reflectors. Unfortunately, the depth of this highly conductive layer and the associated strong reflectors were never reached by wells in the Transdanubian Central Range.

Regional section D is based on the Dd line published by Ádám et al. (1984) between 70 and 120 km (Plate 4). Well #1 reached the Triassic basement beneath Senonian and Neogene strata. At 82 km, well #2 also found Paleogene rocks which can be delineated as a small wedge-like unit pinching out to the NW (Plate 4). The data quality is rather poor beneath the outcropping part of the Mesozoic from about 95 to 120 km. Some reflector packages, however, may be interpreted as thrust units (see Chapter 5).

Probably to the S of the information gap between 120 and 130 km, regional section D crosses the zone of the Balaton Line, since wells #8 and #9 encountered Dinaric-type Lower Permian carbonates (Bérczi-Makk, 1988a). The most important features of this zone are the S-vergent compressional structures in its northern part where the crystalline basement of the Bakony Mts. thrusts on Paleogene and Mesozoic rocks (Dubay, 1962; Balla et al., 1987). Unfortunately, these structures in the pre-Neogene basement are not visible on the available seismic data. To the S the Early and Middle Miocene volcanic chain, however, can be identified on the seismic lines (cf. also regional sections B and C).
Fig. 3.13. Dc seismic profile published by Albu et al. (1983). For location see Fig. 3.1.
In the Igal high (wells #10, 11 and #12) Dinaric-type, slightly metamorphosed Triassic limestones were found (Körössy, 1990). The Neogene strata indicate again a very young folding over the Igal high.

3.1.5 REGIONAL SECTION E

The northwestern half of regional section E between 0 and 50 km is similar to the nearby section D (Fig. 3.1). The Pinnye high at about 15 km (well #1) here consists of two fault blocks bounded by low-angle normal faults. The fault bounding the Csapod Basin on its northwestern flank can be traced into the basement and shows a clear low-angle character. Based on the reflection pattern on the southern side of the Mihályi high near 40 km, the Rába fault can be identified, again as a low-angle normal fault. The post-rift Sarmatian strata are anomalously thick in the center of the basin as proven by a nearby well (well #2).

The deep syn-rift trough in the center of the Danube Basin between 40 and 60 km contains Middle Miocene intermediate volcanites according to boreholes (Pásztori) just to the S of regional section E at 55 km. To the SE the pre-Tertiary basement is made up of unmetamorphosed Triassic carbonates, Permian clastics and underlying very low-grade Paleozoic slates (wells #3 and #5). The basement reveals an internal structure suggesting thrust imbricates from about 50 km to the SE (see Chapter 5). A marked change in the reflection patterns indicates a major thrust plane reaching the surface of the basement at 70 km.

A similar thrust contact is shown on the Ea line (Fig. 3.14), which is located some kilometers to the S of regional section E (Fig. 3.1). Horváth and Rumpler (1984) interpreted the prominent reflectors in the basement as related to a NW-vergent Alpine overthrust plane sealed by the Neogene succession. Note also that the well-developed prograding unit within the Pannonian sequence shows a post-depositional dip caused by
Fig. 3.14. Ea seismic line published by Horváth and Rumpler (1984) with the migrated (a) and line drawing interpretation (b). For location see Fig. 3.1.
the young uplift of the Bakony Mts.

Fig. 3.15 shows the Eb line published by Pápa et al. (1990), which is part of a crustal seismic section to the NE of regional section E (Fig. 3.1). The crustal seismic section can be found in Plate 5. The detail of the intra-basement reflection patterns in Fig. 3.15 was also interpreted by Pápa et al. (1990) in terms of NW-trending thrusts.

3.1.6 REGIONAL SECTION F

The regional section between 0 and 70 km is based on a seismic section published by Dudás et al. (1987). The authors interpreted a number of normal faults dipping to the SE between 0 and 25 km controlling the deepest syn-rift portion of the Danube Basin. On the regional line, however, a major low-angle normal fault is interpreted tentatively on the base of seismic patterns. The approximate top of the pre-Tertiary basement and the Middle Miocene syn-rift is also shown.

Interestingly enough, the whole post-rift sequence displays a slight synclinal sag over the entire width of the basin. It is important to realize that this syncline became evident because of the vertical exaggeration of the seismic data. This feature can be explained only in part by differential compaction and the bulk of the structure may be caused by very young compression (see Chapter 5). Indeed, on the SE flank of the Danube Basin, even the youngest Pannonian reflectors show a near-surface truncation. The same phenomenon can also be seen on the previous regional lines.

Between 70 and 122 km, regional line F is based on the Fa line (Fig. 3.1) published by Posgay et al. (1986) and a part of it is shown in Fig. 3.16. This crustal reflection seismic line can be found in Plate 6 with a new interpretation. The data hardly permit interpretation of steeply dipping normal faults crossing the upper and middle crust as shown in Fig. 3.16. Instead, the pronounced reflection packages within the basement suggest thrust tectonics (Plate 6) comparable to the previous regional sections.
Fig. 3.15. Detail of the Eb deep seismic reflection line published by Pápa et al. (1990). The entire profile can be found in Plate 5. See Fig. 3.1 for location.
Fig. 3.16. Detail of the Fa deep reflection seismic line from Poggey et al. (1986). See Fig. 3.1 for location.
There is a wedge of Upper Paleogene sediments between 67 and 80 km onlapping on the basement (allowing for the reconstruction of the pre-rift geometry). A similar but less pronounced wedge can be seen on the regional section E between 73 and 82.

Fig. 3.17 shows the Fb line published by Albu et al. (1983). This line is located to the SW of regional section F (Fig. 3.1) and corresponds to the portion between 92 and 110 km. On the Fb line, there is again good correlation between the conductive layer and a set of prominent intra-basement reflectors on the NW side of the section.

There is a gap in the regional section between 122 and 140 km in the area of the Velence Mts. (Fig. 3.1). The Fc line (Fig. 3.18) is located along strike, several kilometers to the SW. The Middle Miocene syn-rift strata of the Polgárdi depression experienced compression before the post-rift phase according to Balla and Dudko (1990).

The regional line follows the trace of a crustal seismic section from 140 km to the end of the section, close to the Danube (Fig. 3.1). Basement is apparently at shallow depths, but very poor data quality prevents detailed interpretation.

3.2 EASTERN HUNGARY, DIP LINES

These regional lines (Panel 2) cross the area of the Great Hungarian Plain from the Hungarian Mid-Mountains in the N to the Serbian and Romanian border in the S, striking to the NW-NNW (Fig. 3.1). The ties with the strike lines are also indicated on Panel 2.

3.2.1 REGIONAL SECTION G

At the northern end of the section, between 2 and 10 km, the deep Csepel Trough is bounded by steep faults. There is no well control to date on the fill of this trough, which is tentatively regarded as Miocene. To the S of a basement high, chaotic reflectors beneath the post-rift succession between 17 and 38 km are interpreted as Miocene volcanites. Note that the regional section between 10 and 40 km may be correlated with the zone of Middle
Fig. 3.17. Fb seismic line published by Albu et al. (1983). Resistivities are given in ohm/m. For locality see Fig. 3.1.
Fig. 3.18. Fc seismic line published by Balla and Dudko (1990) with its migrated version (a) and line drawing interpretation (b). For location see Fig. 3.1.
Miocene volcanites shown on regional lines B, C and D (Panel 1). The Mid-Hungarian Line is again poorly defined and may be placed beneath the volcanics (cf. Dank and Fülöp, 1990). In any case the two troughs and the basement high in between may be traced along strike using a filtered gravity map (Szabó, 1989).

Based on nearby wells (see Kerekegyháza basement high on section H) the pre-Tertiary basement is made up of Mecsek-type Mesozoic from about 40 km southward. To the S, wells #1, #2, #3 and #4 indeed found the characteristic Paleozoic granite and Mesozoic carbonates (Körössy, in press).

An interesting structure at about 60 km is best illustrated by the nearby Ga line (Fig. 3.19) published by Pogácsás et al. (1989), who suggested that the local trough in the NW is bounded by a N-dipping listric fault. A more viable alternative was adopted on the regional section, suggesting a S-dipping thrust whose very young activity is responsible for the anticlinal structure above the basement high. Thus a set of strong reflectors beneath well #1 at about 2 s TWT time is regarded as fault plane reflectors (Fig. 3.19). A comparable Pliocene thrust was documented by Wein (1965) in the Northern Thrust Zone of the Mecsek Mts., some 70 km to the SW (Fig. 3.1), where Mesozoic rocks were thrust on Lower Pannonian (Upper Miocene) sediments (cf. Tari, 1992c).

In the Kecel area well #5 encountered one of the mounds on top of the basement shown by the seismic between 80 and 90 km and it found Upper Miocene (Lower Pannonian) basalts and tuffs.

Wells #7, #8, #9 and #10 drilled the medium-grade metamorphics of the Villány unit (Fig. 3.2). The deep syn-rift graben between 100 and 120 km is the Kiskun depression, which is the only documented pull-apart basin in Hungary (P. Régyli, unpublished). This ENE-trending basin is about 20 km wide and 35 km long, bounded by a left-stepping pair of master faults. The Gb line shown in Fig. 3.20 is perpendicular to regional line G at 108 km and is located in the long axis of the basin. According to Rumpler and Horváth
Fig. 3.19. Ga reflection seismic line (Pogácsás et al., 1989). For location see Fig. 3.1.
Fig. 3.20. Gb seismic reflection line (Rumpler and Horváth, 1988) with its stacked version (a) and line drawing interpretation (b). For location see Fig. 3.1.
(1988), not only the Sarmatian but also the lower part of the Pannonian is missing on the erosional top of the thick (>3 km) syn-rift strata (e.g. well #10). This is best explained by an inversionsal folding (see regional line) and uplifting of the basin fill above sea-level before the post-rift phase.

The tectonic boundary between the Villány and Békés units can be placed between wells #11, #12 and #13, #14, respectively, based on the contrasting facies of Mesozoic rocks (see Fig. 3.2b).

3.2.2 REGIONAL SECTION H

At the northern end of the section between 0 and 8 km, a deep graben is correlatable with the one shown in the same position on regional line G. To the S of this Miocene graben, a number of N-dipping reflectors between 10 and 16 km are truncated by the basal Middle Miocene unconformity. The steeply dipping beds are thought to be Paleogene, based on the nearby wells of the Bugyi high. The so-called Paleogene Line which refers to the sharp southern boundary of the Paleogene Basin (Vadász, 1960) is placed just to the S of the Bugyi high (e.g. Csiky, 1961).

Well #1 reached Dinaric-type Permian carbonates below a probably Eocene (or Oligocene?, e.g. Juhász, 1971) volcanoclastic sequence. In fact, two volcanic cones can be seen on the regional section, at 19 and 24 km. Farther to the S, between 26 and 43 km the Örkény Trough is shown filled with Miocene volcanites based on nearby Örkény, Táborfalva and Újszilvás wells. This deep graben (up to 5 km) shows up on the pre-Tertiary basement map of Kilényi et al. (1991) as a pull-apart basin (see also Rumpler and Horváth, 1988) similar to the Kiskun depression described earlier on regional line G. The wide zone of Miocene volcanics here again covers the postulated Mid-Hungarian Line, since the basement on the Bugyi high around 20 km is made up of Bükk-type (North Pannonian) Mesozoic and Paleozoic (Bérczi-Makk, 1978a), while in the Kerekegyháza
high at about 47 km Mecsek-type Mesozoic was drilled (e.g. Fülöp and Dank, 1987; Körössy, in press).

At 70 km, a few kilometers wide zone of very young (Quaternary) faulting with a flower structure geometry is shown. In a zone between 60 and 80 km, several Pannonian reflectors are truncated, suggesting regional uplift of the area to the N. Farther to the S, the pre-Tertiary basement gradually deepens to 5-6 km depth. Several basement highs are shown on the regional line between 90 and 170 km. Note the structural relief (up to 2-3 km) between these highs and the intervening basins. Because of the large relief, syn-rift sediments can be found only in the deeper basinal areas, and the post-rift Sarmatian or the lower Pannonian strata typically onlap on the flanks of the basement culminations. On the Pálmonostora high well #6 encountered crystalline rocks assigned to the Villány unit. Farther to the S wells #7 and #8 drilled Triassic nonmetamorphosed and Paleozoic medium-grade metamorphics of the Békés unit (Fig. 3.2b).

A major southward prograding delta system within the Pannonian succession becomes very pronounced from about 80 km. The geometry of the clinoforms suggests progradation in about 600-800 m water depth (e.g. Mattick et al., 1988). The prograding unit drapes over the basement highs mainly because of the differential compaction of the thick post-rift strata and perhaps partly because of the same young compression period documented in Western Hungary (see regional lines A-F).

3.2.3 REGIONAL SECTION I

This section starts close to the Slovakian border (Fig. 3.1). In this area, shown between 0 and 17 km, the Oligocene and Miocene rocks outcrop on the surface (e.g. Hámor, 1985). At 21 km, however, along a major fault the Oligo-Miocene succession is downthrown into the Zagyva Basin, to the W of the Mátra Mts. The fault itself shows a very young reactivation revealed by folded Pannonian reflectors.
The section runs along the axis of the N-S trending Zagyva Trough between 20 and 45 kms. At about 3-4 km depth the pre-Tertiary basement comprises Bükk-type Mesozoic ophiolites (well #1). The basement is overlain by a 1-2 km thick Paleogene sequence, below the partly volcanic Middle Miocene syn-rift strata. There is no obvious correlation between the thickness and distribution of the Paleogene and the Miocene strata. This suggests the superposition of the Neogene Pannonian Basin on an earlier Paleogene basin of different origin (cf. Fig 2.2).

The next four seismic illustrations are located at the northern end of section I, in the area of the Zagyva graben (Fig. 3.1). Section Ia (Fig. 3.21a) is a dip section, striking E-W across the Zagyva Trough. The structural interpretation (Fig. 3.21b) shows that this deep half-graben is bound to the E by a steeply-dipping master fault, whereas the western gentle flank of the basin is affected by several subordinate antithetic and synthetic normal faults. In the westernmost part of this section, another half-graben can be found, which is also delimited by a west-dipping master fault on its eastern side. The amount of extension across the entire section is about 5-10%. Both master faults can be traced to the N on successive dip lines; however, to the S, seismic profiles with roughly the same orientation show a sudden switch not only in dip polarity of the normal faults but also in the style and amount of extension (section Ib, Fig. 3.22a). The strongly rotated fault blocks (Fig. 3.22b) are separated by listric faults, dipping gently to the E. These faults could also be traced on several successive dip lines. The amount of extension in this section is estimated as high as 20-30%. Farther to the S, section Ic (Fig. 3.23a) shows hardly any extensional structural features (Fig. 3.23b), indicating negligible E-W extension in this area.

Fig. 3.24a gives the strike line Id which is part of the section between about 45 and 65 km (Fig. 3.1). Two fault zones are recognized in this profile (Fig. 3.24b) that resemble flower structures suggesting zones of strike-slip movements. Since these fault zones coincide with the boundaries of the above described extensional areas characterized by
Fig. 3.21. Ia migrated seismic section (a) and its interpretation (b) from Tari et al. (1992b). For location see Fig. 3.1.
Fig. 3.22. Lb migrated seismic section (a) and its interpretation (b) from Tari et al. (1992b). For location see Fig. 3.1.
Fig. 3.23. In migrated seismic section (a) and its interpretation (b) from Tori et al. (1992b). For location see Fig. 3.1.
Fig. 3.24. 1d migrated seismic section (a) and its interpretation (b) from Tari et al. (1992b). For location see Fig. 3.1.
different styles and amounts of extension, Tari et al. (1992) interpreted them as transfer faults sensu Bally (1981) and Gibbs (1984). This interpretation is compatible with the definition of transfer faults as zones of differential lateral motion accommodating contrasting extensional regimes.

The southern transfer fault shows some recent reactivation (see Fig. 3.24b). This particular transfer fault probably continues toward the ENE (see Kb line and regional line K) and correlates with the master fault of the Vatta-Maklárf Trough (Tari, 1988). If this is indeed the case, this single transfer fault can be traced for a distance of more than 100 km (see Fig. 3.2b). Based on newly acquired seismic data to the W of the Zagyva Trough the northern transfer fault is not as well defined as I previously supposed (cf. Fig. 8. of Tari et al., 1992). My preliminary interpretation of this new seismic data (vintage of 1992 and 1993) suggests a wide (5-10 km) transitional accommodation zone rather than a sharp boundary fault between the two extensional fault domains shown in Figs. 3.21 and 3.22.

The Mesozoic basalts drilled in well #2 are comparable to those of the Bükk Mts. The Paleogene sequence can be followed to the S to about 78 km where it abruptly ends against a fault plane that may be identified as the Paleogene Line. Farther to the S, in the area between 80 and 100 km the syn-rift sequence is made up of volcanites penetrated by wells drilled on the nearby Farmos high (e.g. wells #3 and #4). These volcanites may indicate the poorly constrained zone of the Mid-Hungarian Line (cf. regional section H above). The pre-Miocene basement is very poorly known in this zone.

Well #5 found Mesozoic carbonates assigned to the Mecsek unit of Tisia. Between about 115 and 130 km the regional section crosses the westernmost part of the Cretaceous and Paleogene Szolnok flysch belt (well #6, see Fig. 3.2). The flysch does not have any distinct seismic expression and therefore its exact boundaries cannot be determined.

The Ie line (Fig. 3.25) is located some 10 km to the E from well #6 (Fig. 3.1). In this line, the Szolnok flysch is thrust to the N according to Lörinc and Szabó (1992), on top of
Fig. 3.25. 3D reflection seismic line of Lörinc and Szabó (1992). For location see fig. 3.1.
Mecsek-type Liassic rocks encountered in the nearby Abony area (e.g. Fülop and Dank, 1987). Löricz et al. (1989) regarded the internal folding and imbrication of the flysch sequence reported by many (e.g. Szepesházy, 1973; Báldi-Beke and Nagymarosy, 1992) as the result of Late Miocene transpressional movements.

From about 100 km onwards, regional section I shows the deepening of the pre-Tertiary basement to a depth of 4-5 km. In this area several S-dipping normal faults were interpreted as bounding small syn-rift grabens. Taking into account the vertically exaggerated scale of the regional lines, these faults are in fact low-angle normal faults, indicating significant amounts of extension. Some of these low-angle normal faults are tentatively interpreted as reactivated Eoalpine overthrusts, as for example the supposed contact between the Villány and Békés units at about 170 km.

Note the pronounced change in the geometry of the Pannonian prograding unit at about 140 km shown as a sequence boundary. The abrupt flattening of the clinoforms suggests a sudden decrease of water depth from about 800 m to 200 m (see later).

3.2.4 REGIONAL SECTION J

The regional section starts near the outcropping Miocene volcanites of the Mátra Mts (Fig. 3.1). At about 10 km the N-dipping normal fault may be correlated with the southern transfer fault of Fig. 3.24 on regional line I. Farther to the S, the syn-rift Miocene consists of dominantly volcanites according to well #1. The Ja line of Kílényi et al. (1991) shown in Fig. 3.26 is located some 15 km to the E (Fig. 3.1). In this latter section, a thick Miocene volcanic sequence was also drilled on top of the Bükk-type Triassic in the Kömlő-1 well. The deep basin between about 40 and 60 km is probably filled by volcanites, indicating the zone of the Mid-Hungarian Line.

The deep seismic Jb line published by Posgay and Szentgyörgyi (1991) runs parallel to the regional line from about 60 km southward and some 20 km to the E (Fig. 3.1). A
Fig. 3.26. Ja seismic profile (Kilényi et al., 1991). For location see Fig. 3.1.
part of the Jb line named as the Pannonian Geotraverse by Posgay and Szentgyőrgyi (1991) is shown in Fig. 3.27, while the original data and an alternative interpretation are enclosed in Plate 7. The authors identified the Mid-Hungarian Line as a zone of disturbance between 74 and 81 km of their section shown in Fig. 3.27. Moreover, this fault zone crosscuts the whole crust and can be followed also into the mantle according to Posgay and Szentgyőrgyi (1991).

The Szolnok flysch was drilled on the Nagykörút high in several wells (e.g. wells #3, #4 and #5). The extension of this flysch to the S cannot be determined from the available seismic data alone.

Farther to the S, similar to regional section I, S-dipping low-angle normal faults appear from about 100 km as the regional line enters the deeper part of the Great Hungarian Plain. As the regional line J represents a dip line running perpendicular to the direction of Neogene extension (cf. regional lines P, Q and R), the low-angle character of the normal faults suggests considerable NNW-SSE extension along the section.

From well #6 Pap (1990) reported an Eoalpine(?!) overthrust, in which an approximate 300 m thick Jurassic sequence is lying on Lower Cretaceous basalts. This intra-basement deformation, however, does not have any definite seismic expression.

The regional line coincides with the Jc line shown in Fig. 3.28 between 144 and 155 km (Fig. 3.1). Note that the upper 1 s of the seismic section was omitted by Grow et al. (1989). On the Békés high in well #11 a more than 1000 m thick Mesozoic succession was drilled on top of a Senonian-Triassic section. In this area the low-frequency intra-basement reflectors in Fig. 3.28 may correspond to a thrust plane. Based on many well-documented cases of Eoalpine thrusting in the pre-Tertiary basement of SE Hungary (e.g. Pap, 1990) and correlating them using seismic data, Grow et al. (1989) outlined several N-verging nappes within the Békés unit correlatable with the outcropping Codru nappes of the Apuseni Mts. (Fig. 3.1). The authors also proposed a probable Miocene reactivation of the
Fig. 3.27. Detail of Jb deep reflection seismic line published by Posgay and Szentgyörgyi, (1991). For location see Fig. 3.1. For the entire seismic section see Plate 7.
Fig. 3.28. 3C seismic profile of Grow et al. (1989). For location see Fig. 3.1. The low-frequency intra-basement reflectors correspond to an Alpine thrust plane found in well #2.
Cretaceous overthrusts as low-angle normal faults.

The depth of the pre-Tertiary basement in the Békés subbasin may exceed 5-6 km. At the end of the section, on the Pusztaföldvár high several wells encountered high-grade metamorphics (wells #14 and #15) with stripes of unmetamorphosed Triassic in between (wells #13 and #16; see also the monography of Kovács and Kurucz, 1984).

3.2.5 REGIONAL SECTION K

The Ka line shown in Fig. 3.29 is located some 40 km to the N of the northern end of the regional section (Fig. 3.1). The high-frequency seismic section was published by Braun et al. (1989). Their original interpretation and an alternative by Sztanó and Tari (1993) can also be found in Plate 8. The Ka line crosses the Darnó Line on the northwestern flank of the Bükk Mts. This NE-trending enigmatic fault zone has a complicated transpressional history based on microtectonic and surface relations (e.g. Zelenka et al., 1983; Csontos, 1988). The inferred strike-slip component cannot be resolved from the seismic data, but the Ka line suggests a Late Oligocene period of NW-vergent thrusting. So far, this is the only seismic evidence for Paleogene tectonics in the Pannonian Basin. The folding of Lower Miocene strata on the Ka line indicates another, less pronounced post-Early Miocene episode of compression. Alternatively, this structure may be interpreted as a triangle zone sensu Jones (1982).

At the northern end of regional line K, the Vatta-maklár Trough can be found between 0 and 10 km. The Kb line published by Pogácsás (1984) illustrates the structure of this deep half-graben some 2 km to the W (Fig. 3.30). The base of the Miocene syn-rift sequence corresponds to the unconformity between Units 1 and 2 of Pogácsás (1984). The deeper part of the basin is filled by thick (>1 km) Paleogene strata (Tari, 1987, 1988).

On the basement high of Mezőkövesd, between 10 and 20 km, Miocene volcanic strata were drilled in wells #1, #2 and #3 overlying an erosionally truncated Paleogene
Fig. 3.29. Ka migrated seismic section (a) and its line drawing interpretation (b) from Braun et al. (1987) and Sztanó and Tari (1993), respectively. For location see Fig. 3.1. See also Plate 8.
Fig. 3.30. Kb seismic profile (Pogácsás, 1984). For location see Fig. 3.1. Legend: 1) pre-Neogene (Paleogene and Mesozoic), 2-6) pre-Pannonian Miocene, 7-9) Pannonian strata.
sequence. The pre-Tertiary basement is made up of Bükk-type Triassic carbonates (Bérczi-Makk, 1978b). Well #4 bottomed in Eocene marls (Körössy, in press); therefore the enigmatic Paleogene Line should be placed to the S of it. Between about 20 and 50 km, the deep Miocene volcanic trough (well #5) of the Mid-Hungarian Line can be found.

At around 60 km in the Kunmadaras area, the Szolnok flysch was found in several wells (e.g. #9). Farther to the S on the Karcag-Bucsa basement high at 82 km, the flysch was drilled again (wells #10 and #11). In this region the flysch is characterized by many strong and generally N-dipping reflectors. The basement underlying the flysch is not known. Well #12 encountered medium-grade metamorphics. On this Füzesgyarmat high Mesozoic rocks assigned to the Villány zone were also found.

From about 85 km, the regional section crosses the deeper part of the Great Hungarian Plain. The syn-rift sequence is thin in subbasins and missing on basement highs. The deepest part (5-6 km) of the area is located to the W of the Komádi basement culmination between 130 and 135 km. This half-graben is controlled by a W-dipping low-angle normal fault; however, this low-angle character is mainly due to the slightly oblique intersection with the fault (cf. regional section Q).

3.2.6 REGIONAL SECTION L

Regional section L crosses a deep Miocene half-graben at its northern end between 3 and 15 km. The pre-Tertiary basement in this area and farther to the NE is poorly known (e.g. Kilényi et al., 1989) because of a thick Middle Miocene volcanic blanket (wells #1 and #2). Therefore the top of the pre-rift succession is shown only tentatively. From about 55 km to the S, the Miocene syn-rift sequence is thinner on the Józsa high and it is underlain by the Szolnok flysch based on wells #4 and #5. In this region the flysch has a seismic expression and shows perhaps some internal folding. In any case the truncation of the flysch beds (between 65 and 75 km) indicates an erosional southern boundary of the

Well #6 drilled almost 1900 m into the pre-Tertiary basement. In this well Pap (1990) documented an about 850 m thick thrust sheet made up of Paleozoic (?) medium-grade metamorphics overlying very low-grade Triassic clastics. This intra-basement overthrusting may have a seismic expression in the form of many N-dipping reflectors (see also Rumpler and Horváth, 1988).

The deep (6-7 km) Derecske Trough between 82 and 95 km is filled by a thick Miocene succession (well #13) suggesting the syn-rift activity of the master fault. The same fault shows very young (Quaternary-Recent?) reactivation forming a flower structure in the post-rift sedimentary fill. The strike-slip character of a similar nearby fault is best illustrated by the La line (Fig. 3.31) published by Kókay and Pogácsás (1991). This line is located some 15 km to the SW of the regional line (Fig. 3.1). The fault has a clear offset in the basement and the accompanying basin was interpreted as the Derecske pull-apart basin by Marton (1985).

3.3 EASTERN HUNGARY, STRIKE LINES

These regional lines (Panel 3) cross the area of the Great Hungarian Plain from the Danube in the W to the Romanian border in the E striking to the NE-ENE (Fig. 3.1). The ties with the dip lines of Panel 2 are also indicated on Panel 3.

3.3.1 REGIONAL SECTION M

The regional line starts from the Danube in the W (Fig. 3.1). From 0 to about 50 km, the section runs very close to the transfer fault discussed earlier on regional lines H and I (Panel 2). Therefore it is difficult to interpret the internal structure of the pre-Pannonian strata. There is no well control in this area but the Paleogene seems to be thick (up to 2 km) and it is covered by a Miocene volcanic syn-rift sequence. At 60 km, the section
Fig. 3.31. La reflection seismic section (a) and its interpretation (b) by Kókay and Pogácsás (1991). See Fig. 3.1 for location.
shows a Paleogene thrust, but similarly to the example shown in Fig. 3.29 this interpretation is also debatable. Note that in order to show an uninterrupted regional section, the orientation changes considerably between 60 and 88 km (Fig. 3.1). Because of this the NE-trending transfer fault described earlier (see Fig. 3.25) is crossed at 80 km.

Between 135 and 150 km, the regional section coincides with the Ma line published by Kókay and Pogácsás (1991) shown in Fig. 3.32. This line runs close to the axis of the Vatta-Maklár Trough shown earlier on regional section K (Fig. 3.1). The graben is underlain by a low-angle normal fault, and in the middle of the section the faults also have normal separation in contrast to the interpreted reverse faults. The W-dipping reflectors shown as collapse graben reverse faults at the eastern end of the section are related to a side-swipe effect (Tari, 1987). These crosscutting out-of-plane reflectors originated from the nearby Mezőkövesd basement high (see regional section K).

On the Mezőkeresztes basement high (wells #5, #6, #7 and #8) the Paleogene is still present but with an erosionally reduced thickness (Bérzi-Makk, 1975). From 175 km to the E, the regional line obliquely crosses the Vatta-Maklár Trough (well #9), giving a misleading low-angle character to the steeply N-dipping transfer fault.

3.3.2 REGIONAL SECTION N

This regional line crosses the Kecel-Lajosmizse-Táborfalva high between 0 and 40 km, where the pre-Pannonian basement is at relatively shallow depth (up to 1 km). The Mesozoic sequence drilled in wells #1, #3 and #4 is made up of Mecsek-type Cretaceous rocks (e.g. Fülöp and Dank, 1987). Well #2 encountered Upper Miocene (Lower Pannonian) basalts (see also regional section H). The shallow depth of the basement is partly the result of Quaternary uplift suggested by E-dipping truncated Pannonian reflectors between 35 and 85 km.

From 40 km to the E, the regional section obliquely enters the deep zone of the mid-
Hungarian Line filled with Early and Middle Miocene volcanics. Wells #5, #6, #7, #8, #9 and #10 all reached these andesites, rhyolites and their tuffs. From the deepest part of the basin at 120 km the volcanic basement becomes gradually shallower again to the E. From about 170 km to the E, the volcanics are probably underlain by the Szolnok flysch although the seismic data do not resolve this change.

Note that in contrast to all the dip sections (A-L) that showed a uniformly southward prograding Pannonian delta complex in this strike section and the following sections, two directions of progradation can be observed. In regional line N, the two delta fronts prograding towards the center of the section join each other at about 120 km. Within the prograding unit a number of sequence boundaries were interpreted (see the description of regional line Q for details). The fault-controlled down-slope slumping at 140 km (see also regional line K) is a unique feature in the Pannonian Basin.

3.3.3 REGIONAL SECTION O

The regional line starts from the middle of the Danube-Tisza interfluve (Fig. 3.1). The pre-Tertiary basement is made up of Mecsek-type Cretaceous magmatites and carbonates according to wells #1 and #2 near Kecskemét. From about 35 km, the pre-Neogene basement consists of Szolnok flysch according to Fülöp and Dank (1987). The seismic data record only a slight change, i.e. the basement becomes more reflective to the E.

The deepest part of the basin along this section is located at 61 km, controlled by a syn-rift half-graben. Note that this area coincides with the point where the two facing Pannonian prograding units join each other (cf. also regional section N).

At 85 km, the Kisújszállás basement high is made up of the Szolnok flysch based on numerous wells (e.g. #3, #4, #5 and #6). Farther to the E, from about 140 km, the flysch (#7 and #8) is overlain by Middle Miocene volcanics on the Kaba high (well #9). The
exact boundary is hard to draw judging only from seismic character. The Miocene volcanics show a definite thickening farther to the E (Kilényi et al., 1989). The pre-Tertiary basement in the whole eastern half of the section is unknown since not a single well penetrated the volcanics and/or the flysch (e.g. Szepesházy, 1973).

In the continuation of regional line O, the Oa line (Fig. 3.1) shown in Fig. 3.33 illustrates the seismic expression of the thick Miocene volcanic complex (Kilényi et al., 1989). The Nagyecsed-1 well first penetrated rhyolitic-andesitic tuffs below the Pannonian sequence. Andesitic lavas were found at deeper levels, and at the bottom granodiorite was found (Fig. 3.33).

3.3.4 REGIONAL SECTION P

In contrast to the previous strike lines (M, N and O), the next three lines (P, Q and R) are located in the highly extended and deepest region of the Great Hungarian Plain in SE Hungary. Regional line P starts at the Serbian border. The Ásotthalom basement high between 0 and 10 km is separated by an E-dipping low-angle normal fault from the Algyö high between 15 and 30 km. E-dipping low-angle normal faults are common in this area as was suggested by Rumpler and Horváth (1988). Their seismic example is shown in Fig. 3.34 as the Pa line (Fig. 3.1). In the Üllés and Forráskút area the abundant well control confirmed the presence of these faults (e.g. wells #1, #2, #3 and #4).

Between 30 and 80 km, the Makó Trough is shown as the deepest subbasin in SE Hungary (7-8 km). A slight asymmetry in the syn-rift basin fill indicates that the basin was formed by low-angle normal faulting on its western flank (see regional lines Q and R below). Miocene strata onlap on the western flank of the broad Pusztaföldvár basement high between 60 and 85 km (well #5). The lowermost part of the post-rift strata, including the Sarmatian and the lower Pannonian, also onlap on the basement. At 80 km, a set of strong intra-basement reflectors suggests the presence of Cretaceous nappes.
Fig. 3.33. Oa seismic line with the interpretation of Kilényi et al. (1990). For location see Fig. 3.1.
Fig. 3.34. Pa stacked seismic line (upper) and its line drawing interpretation (lower) by Rumpler and Horváth (1988). For location see Fig. 3.1.
(cf. Fig. 3.28).

Between 120 and 170 km, the E-dipping low-angle normal faults are replaced with a series of steeper and W-dipping normal faults. From 170 km, to the NE, the regional line obliquely enters the Derecske Trough discussed previously on line L. The curved faults dipping to the E around 200 km are the extension of the flower structure shown in regional section L. The Szolnok flysch displays strong internal reflectivity. A band of subhorizontal reflectors at about 3 s TWT time may be interpreted as the base of the flysch, but more probably it corresponds to a regional detachment of Eoalpine(?) thrusts documented in well #13 (same as well #6 of regional section L; Pap, 1990). As to the base of the Szolnok flysch, at the bottom of the deep well #14 Lower Cretaceous(?) rocks were described with a lot of uncertainty.

In this regional line, only the northeastern Pannonian prograding delta system can be observed. The southwestward prograding delta shows the characteristic change from a deep water (600-800 m) to a shallow water (200-300 m) environment (e.g. Mattick et al., 1988) at about 175 km.

3.3.5 REGIONAL SECTION Q

The southwestern end of the regional line between 0 and 23 km coincides with the Qa profile (Mattick et al., 1988) shown in Fig. 3.35. The Algyö crystalline basement high was reached by several boreholes (e.g. wells #1 and #2). In the center of the Makó trough, well #3 (deepest in Hungary) bottomed in the upper part of the Middle Miocene syn-rift strata (Badenian) at 5876 m (Bérczi, 1988). Note that the detachment fault on the western flank of the Algyö high is suggested by the subtle asymmetry of the basin fill, especially by a gentle inversional anticline (Fig. 3.35). This interpretation is reconfirmed by a recent, still unpublished crustal seismic section (PGT-3, vintage 1993) which clearly shows the E-dipping detachment fault beneath the Makó subbasin.
The Pusztaföldvár crystalline basement high was reached between 40 and 55 km by wells #4, 5, 6 and #7. Farther to the NE from about 80 km, the extension was accommodated by a number of W-dipping normal faults that might be antithetic to another low-angle fault on the eastern flank of the Pusztaföldvár high. This detachment fault is responsible for the formation of the Békés Basin becoming deeper to the SE of the regional section. The W-dipping steeper normal faults are correlatable with those interpreted in regional line P, and between about 70 and 110 km they bound blocks made up of Mesozoic units (wells #8 and #9).

The prograding Pannonian delta unit between 135 and 158 km is best illustrated by the nearby Qb seismic line (Fig. 3.1) shown in Fig. 3.36a. Based on stratigraphic patterns, (e.g. Vail, 1987) several depositional sequences can be recognized in the section (Fig. 3.36b). All of them are characterized by erosion on the delta plain and pronounced downlap and toplap at the base and at the top of the highstand systems tracts, respectively. The lowstand systems tracts are not fully developed except for the sequence on the southwestern part of the section. The sand sheet of the basin floor fan and the overlying slope fan mounds indicate a major fall in the water level of the Pannonian lake perhaps associated with the Messinian event in the Mediterranean (see Appendix B for details).

Taking into account that during this time interval the progradation of the Pannonian delta system was extremely rapid, about 30 km/Ma (e.g. Pogácsás et al., 1988), the average duration of each sequence in Fig. 3.34 is estimated as 0.1-0.5 Ma. Therefore they exemplify fourth-order sequences (sensu Van Wagoner et al., 1989) or subsequences (Carlos Cramez, pers. comm.).

The depositional sequences associated with basin floor and slope fans (Fig. 3.34b), however, indicate a major fall in lake level that can be mapped regionally (Matticek et al., 1988). This "destructive phase" (Matticek et al., 1988) separates earlier deltas prograding into a 800-900 m deep lake from a subsequent delta system characterized by progradation
Fig. 3.36. Qb migrated seismic section (a) and its interpretation (b) by Tari et al. (1992b). For location see Fig. 3.1.
in shallower water (200-400 m). In fact, this particular event represents a third-order sequence boundary (Tari et al., 1992b). Since Pogácsás et al. (1988) established a correlation between similar unconformities in the nonmarine sedimentary fill of the Pannonian Basin and the eustatic curve of Haq et al. (1987), a question arises concerning the reason for this coincidence. The two alternative explanations offered by Tari et al. (1992a,b) are discussed in detail in Appendix B.

The Qc line shown in Fig. 3.37a is located some 10 km to the SE of regional line Q, close to the Romanian border (Fig. 3.1). This seismic section published by Tari et al. (1992b) illustrates syn-rift low-angle normal faulting. In the SW, numerous wells reached the basement (Biharkeresztes high) at about 2 km depth. Here the basement is made up of Early Paleozoic gneisses. At the northeast end of the section, not far from this line, a drillhole penetrated the same type of high-grade metamorphic basement. However, in a borehole in between (Nagykereki-1) thick Middle Miocene strata were found. Interpretation of the section (Fig. 3.37b) shows that this Miocene succession has a prominent tilt into a low-angle fault plane, that is suggested by faint fault plane reflections. This fault has a listric geometry and the tilted Miocene reflectors give evidence for syndepositional growth. Correlation of the basal Miocene succession with a narrow Miocene trough, trending perpendicular to section at the northwest side of the Biharkeresztes basement high, indicates approximately 20 km of horizontal offset along this single low-angle normal fault.

3.3.6 REGIONAL SECTION R

This line shows the crystalline Algyő basement high between 5 and 20 km (wells #1 and #2). Here the Makó Basin is controlled by the same low-angle normal fault as on regional line Q. A package of intra-basement reflectors beneath this high suggests its tilting to the SW due to normal faulting (cf. Fig. 3.36). Note that because of the change in
the orientation of the regional line there is a break in the fault plane between 22 and 30 km. On the flank of the Algyő high well #3 encountered the down-faulted Mesozoic rocks of the Békés zone. In the Makó Basin itself the lower part of the post-rift sequence onlaps the Algyő and the Pusztaföldvár (wells #4, #5, #6, #7 and #8) basement highs on the SW and on the NE, respectively.

Farther to the SE of the Pusztaföldvár high, the deep (up to 7-8 km) Békés Basin is located. This subbasin is also flanked by a low-angle normal fault on its southwestern side. The Ra line (Fig. 3.38) published by Mattick et al. (1988) is located to the E of the regional line (Fig. 3.1). In this area, the detachment faulting resulted in a relief difference of about 4 km between the Pusztaföldvár basement high and the adjacent Békés Basin. This Middle Miocene syn-rift topography disappeared only some 10-12 Ma later, when the progressively onlapping post-rift sediments filled the basin.

The Pannonian strata above the Pusztaföldvár high show a broad anticline. Since even the youngest beds are involved in the anticline, this feature cannot be entirely attributed to differential compaction. A very young (Quaternary?) compressional episode might be responsible for this feature in the same way as regional sections A-F suggested in W-Hungary.

The Rb line (Fig. 3.39) published by Dank (1988) shows the Sarkadkeresztúr basement high located to the NE of the regional line, close to the Rumanian border (Fig. 3.1). This swell, made up of medium-grade metamorphics, has a much greater height/width ratio than that of the surrounding basement highs. Therefore some faults developed right on top of the swell in the post-rift succession to accommodate the stresses caused by differential compaction. A potential pitfall of these "draped" faults is that they could be interpreted in terms of strike-slip related flower structures.
Fig. 3.38: Ra reflection seismic line from Matick et al. (1988). For location see Fig. 3.1.
Fig. 3.39. Rb seismic line across the Sarkadkeresztúr basement high (Dank, 1988). For location see Fig. 3.1. The two concave-upward faults above this basement in the post-rift succession swell could be interpreted in terms of strike-slip related flower structures. However, these faults more likely developed right on top of the swell to accommodate the stresses caused by differential compaction.
3.4 CHAPTER DISCUSSION AND CONCLUSIONS

Several conclusions can be drawn from the above described 18 regional profiles and the accompanying 35 seismic illustrations. Many of these conclusions do not follow some commonly accepted views; others are regarded as new points which emerged from this compilation of regional seismic data.

The currently available seismic data clearly do not permit a detailed study of the pre-rift basement of the Pannonian Basin. This is because during the last decades industry efforts were concentrated on the Neogene basin fill. Thus routine acquisition and processing of seismic data were adjusted to this need. Even the thick Paleogene basin fill of N-Hungary is not resolved sufficiently to reveal its internal structure. Many of the major structural lines crossing Hungary (Fig. 3.2), that were postulated on the basis of outcrop and well data, do not have a clear seismic expression (e.g. Mid-Hungarian Line). An exceptional area is the Hungarian Danube Basin where pre-rift structures and basins can be resolved and mapped to a certain degree using conventional seismic data combined with well information (see Chapter 5).

Looking at the Neogene basin fill of the Pannonian Basin at the scale of the presented regional lines, it is clear that the syn-rift sequence is typically thin and contrasts with the thick post-rift succession. This phenomenon was attributed by many (e.g. Royden, 1988) to the starved character of the syn-rift basins caused by the considerable distance from the surrounding Carpathian thrust-fold belt as the source of sediment supply. The thin basin fill makes the interpretation of syn-rift structures very difficult due to the lack of detailed data that may suggest growth associated with faulting. Thick syn-rift successions are found locally only in basins that might be related to a pull-apart mechanism. However, very few documented examples of pull-apart basins exist in the area and many others interpreted as such are better understood in terms of simple extension (e.g. Danube Basin).
It seems that the role of Miocene strike-slip faulting was overemphasized during the last decade (e.g. Tari, 1988). Further confusion arose from the fact that numerous flower structures were observed within the post-rift sequence throughout the Pannonian Basin (e.g. Pogácsás et al., 1989). On one hand, not all of them are related to strike-slip faulting (see e.g. Fig. 3.39). On the other hand, these structures are related to very recent, and in some cases currently active, fault zones (e.g. Fig. 3.31) and cannot be easily assigned to the Middle Miocene syn-rift stage of the Pannonian Basin. Instead, they suggest the neotectonic inversion of the basin (see below).

The regional sections also suggest that the Miocene extension is distributed in a heterogeneous manner throughout the basin. The reflection seismic data from different subbasins of the Pannonian basin complex show very different styles of upper crustal extension. Some areas appear to be affected only by steeply dipping planar normal faults, as in N-Hungary (e.g. regional line I). Seismic sections such as the Ia line (Fig. 3.21) indicate that extension was less than 20%. Other areas, which are characterized by significantly rotated fault blocks (e.g. Ib line, Fig. 3.22), were extended by 20% to 40%. Finally, the Danube Basin and the southern part of the Great Hungarian Plain are seemingly characterized by up to 200% extension (e.g. Aa, Ba, Pa, Qa and Qc lines, Figs. 3.5, 3.7, 3.34, 3.35 and 3.37, respectively). These two basins were extended in NW-SE and ENE-WSW directions, respectively, along major detachment normal faults. The fact that the Neogene to Recent basin fill is the thickest in these two areas (up to 8500 m) is in accordance with their extremely extended crust.

A closely related problem is whether the low-angle normal faults are reactivated pre-existing thrust faults. This phenomenon was first demonstrated by Bally et al. (1966) in the Canadian Rocky Mountains using seismic data. Since then, many other examples of extensional reactivation of earlier thrusts were found (e.g. Ratcliffe et al., 1986). This mechanism was already inferred in some cases in the Pannonian Basin (e.g. Grow et al.,
1989). Tari et al. (1992) also speculated that the tectonic "pre-conditioning" of the Alpine basement is a key element in the localization of not only normal detachment faults but also in the spatial distribution of strike-slip fault zones. This inferred relationship, however, has never been documented in the Pannonian Basin. Chapter 5 is a first attempt to demonstrate the interplay between the Eoalpine thrusts of the basement and the Neogene extensional detachment faults.

The magnitude of extension in the subbasins of the Pannonian basin system is not constrained by structural studies but only estimated by subsidence and thermal data (e.g. Royden and Dövényi, 1988). Since the wells used in these studies were mainly located on basement highs, the amount of crustal extension is probably underestimated (cf. Sawyer, 1986). An alternative approach to estimating the extension values would be systematic mapping of fault offsets in map view by reflection seismic data. This approach, however, would require better knowledge of the intrabasement structural markers, which is not available for most of the Pannonian Basin. This is especially true in NE-Hungary, where a thick volcanic blanket hides the pre- and also syn-extensional structures.

The geometry of the extensional faults along the regional sections, however, suggests a significant amount of E-W extension, definitely exceeding 100 km across the whole Pannonian Basin (cf. Royden et al., 1983). Moreover, as numerous low-angle normal faults indicate, on the southern half of dip lines (H, I and J), there is a significant amount of N-S extension as well. The pronounced upper crustal extension is commonly ignored in the palinspastic reconstructions of the intra-Carpathian area (e.g. Balla, 1984), even though it should have an important effect on the restoration of the pre-rift paleogeography (see Chapter 6).

The heterogeneous distribution of extension observed in the regional lines has an important implication. Many of the faults which separate areas characterized by different magnitude, polarity, and direction of extension had to be connected by complex transfer
systems. The existence of such faults was already postulated by Horváth and Royden (1981). The first documented transfer fault in the Pannonian Basin is the one that can be traced on the northern part of regional lines G-K (cf. Tari, 1988; Tari et al., 1992b). Another major NE-trending transfer zone is located approximately between regional lines O and P (Fig. 3.2) separating the highly extended zone of SE-Hungary (e.g. the Makó and Békés Basins) from the northern areas which display much less extension (e.g. Szolnok flysch area).

Finally, an important new feature revealed by vertically exaggerated regional lines is the very young, in many cases ongoing, uplift of the basement from below the Neogene succession. This upwarping occurs on different wavelengths. The small features includes local folding and/or thrusting of the post-tectonic cover with the inversion of syn-rift structures (e.g. Ab and Bc lines in Figs. 3.6 and 3.9, resp.). There are, however, large anticlinal features responsible for the characteristic "inselberg"-pattern of present-day outcrops (e.g. Aa and Ra lines in Figs. 3.5 and 3.38, resp.). These two end-member neotectonic features were compiled in Fig. 3.40, largely based on the above described regional profiles.

The map-view suggests that the inversion and uplift are propagating into the intra-Carpathian region from the W (Fig. 3.40). The Quaternary to Recent flower structures are now thought to be related to this new, neotectonic stage in the Alpine evolution of the Pannonian Basin. A possible geodynamic explanation for the driving mechanism of the neotectonic inversion is offered in Chapter 8.
CHAPTER 4
DETAILED PHANEROZOIC STRATIGRAPHY AND TECTONICS OF THE HUNGARIAN PART OF THE NW PANNONIAN BASIN

This chapter is a stratigraphic description of Phanerozoic formations of the NW Pannonian Basin (Fig. 4.1). In contrast to Chapter 2, the pre-Alpine (pre-Permian) stratigraphy is also discussed here, since this succession constitutes the "basement" of the Alpine tectonic units. The Permian-Recent (Alpine) stratigraphic column is subdivided into four major units, corresponding roughly to the Permian-Jurassic (early Alpine), Paleoa Alpine or Cretaceous-Paleocene (Eo Alpine), Eocene-Oligocene (Meso Alpine) and Miocene-Recent (Neo Alpine) tectonic stages. Note that the time intervals covered by these stages are slightly different from those defined by Trümpy (1973, 1980).

Within these major time intervals, a further tectonic and spatial subdivision is made. The formations of different units will be described from bottom to top in the Alpine structural edifice (Fig. 4.2). In this chapter the stratigraphic description and brief inventory of previously proposed or currently held tectonic models of different workers cover only the actual study area (Little Hungarian Plain and the Transdanubian Central Range, Fig. 4.3). Other regions located along strike are described in Appendix A. These latter areas include the eastern end of the Eastern Alps with the Styrian Basin, the western end of the Western Carpathians, and the junction area in between with the Vienna Basin (Fig. 4.1).

The legend of stratigraphic columns is the same as in Chapter 2, shown in Fig. 2.7.

4.1 PRE-ALPINE (PRE-PERMIAN) STRATIGRAPHY OF THE NW PANNONIAN BASIN

Since the pre-Alpine basement of the Penninic units is not known in the transitional zone between the Alps and the Carpathians (Fig. 4.1), only the Austro Alpine nappe
Fig. 4.1. Major tectonic units of the Alps-Carpathians-Pannonian Basin junction area.
Fig. 4.2. Simplified Alpine tectonostratigraphy in the NW Pannonian Basin.
Fig. 4.3a. Depth of pre-Tertiary basement in the NW Pannonian Basin. For location see Fig. 4.1.
Fig. 4.3b. Subcrop of pre-Tertiary basement in the NW Pannonian Basin. For location see Fig. 4.1.
complexes are discussed in this subchapter.

4.1.1 LOWER AND MIDDLE Austroalpine UNITS

Lower Austroalpine units comprise a relatively large area of the polymetamorphic basement in the junction area between the Alps and the Carpathians (Fig. 4.1). The distinct facies of the Permo-Mesozoic cover indicates a closer affinity to the Western Carpathians (e.g. Tatric of the Little Carpathians) than to the Alps (see Appendix A). The isolated outcrops of the Lower Austroalpine in the Sopron and Fertőrákos areas of Hungary (Fig. 4.3) are the direct continuation of the Eastern Alpine realm. They are the outcropping margin of the basement of the Danube Basin in Hungary. To the S, in the area of the South Burgenland Swell, the Lower Austroalpine is either very thin or missing.

The basement nappe of the Middle Austroalpine unit dominates the Central Alps to the SW of the Alps-Carpathians transition zone. However, this unit is also reduced in thickness or missing under the western part of the Danube basin. The unit reappears as the Veporides in the inner Western Carpathians, in the easternmost part of the area shown in Figs 4.1 and 4.3.

The Lower Austroalpine succession in the Hungarian part of the Danube Basin (Fig. 4.4) was subdivided into two major formation groups by Fülöp (1990). The Sopron Crystalline Schist formation group (Sopron Micaschist and Sopron Gneiss) with its characteristic "Grobgneis" (see later) development is in an upper Lower Austroalpine position. The Fertőrákos Metamorphics (Fertőrákos Amphiboliteschist and Fertőrákos Micaschist) are a structurally deeper unit, correlatable with the Wechsel complex of the Lower Austroalpine (see later). The Middle Austroalpine unit is not described in the Hungarian part of the NW Pannonian Basin (cf. Chapter 5).

The Sopron Micaschist outcrops around the town of Sopron. It is also known from a number of boreholes in the Sopron area and farther to the E in the Little Hungarian
Fig. 4.4. The Penninic and Lower Austroalpine formations of NW Pannonian Basin, in territory of Hungary. Based on various sources listed in the text.
Plain, in boreholes Rajka(Raj)-1; Mihályi(M)-4, Mosonszentjános(Mos)-1, 2 and Mosonszolnok(Ms)-2 (Fig. 4.3). The 400-500 m thick formation consists mainly of an andalusite-sillimanite-biotite schist, characteristic for the boundary of the amphibolite and the greenschist metamorphic facies (Hercynian?). The subsequent Eoalpine retrograde metamorphism reached the greenschist phase (Lelkes-Felvári et al., 1982, 1983; Kishází and Ivancsics, 1985). An additional and very characteristic lithology is the so-called "white schist", a leuchtenbergite-bearing muscovite-chloritoid quartzite. The white schists are distinctly associated with low-angle fault zones regarded as Alpine overthrust planes (e.g. Fülöp, 1990). The dominant lithology of the Sopron Gneiss is a locally mylonitic muscovite gneiss. The 300 m thick formation shows amphibolite-phase Hercynian metamorphism (Lelkes-Felvári et al., 1982) and is overthrust by the overlying Sopron Micaschist.

The Fertőrákos Amphiboliteschist outcrops only to the N of the village of Fertőrákos, near the Austrian-Hungarian border (Fig. 4.3). During exploration for thorium, the more than 600 m thick formation was encountered in a number of boreholes. Its lithology is dominated by a garnet-bearing albite-actinolite schist (Hercynian amphibolite facies?) overprinted by a Hercynian or Alpine greenschist phase (Fig. 4.4). The original lithology was probably a mafic volcanic rock of Lower Paleozoic age. The Fertőrákos Micaschist outcrops in the same area, and in the Little Hungarian Plain the Csapod(Csa)-1 borehole (Fig. 4.3) penetrated a paragneiss which is assigned to this formation by Fülöp (1990). The 1000 m thick formation consists of a feldspar-bearing micaschist. Marble layers were also found.

4.1.2 UPPER AUSTROALPINE UNITS

The Upper Austroalpine nappes cover a large portion of the junction area between the Alps and the Carpathians (Fig. 4.1). In the Eastern Alps the Gurktal Nappe complex,
the Graz Paleozoic and the Graywacke zone are composed of low-grade pre-Alpine rock series. The same succession is found as the basement of the Styrian Basin and the central and eastern parts of the Hungarian Danube Basin. An important area of outcropping very low to low grade Paleozoic rocks is located in the southeastern flank of the Transdanubian Central Range, in the Balaton Highland and Velence Mts. (Fig. 4.3).

The Graywacke Zone of the Eastern Alps can be traced below the Vienna Basin. It is missing by erosion farther to the NE. The Upper Austroalpine nappes of the Northern Calcareous Alps and in the Western Carpathians (Choc, Lunz and "higher" nappes) have a Late Paleozoic base.

The pre-upper Permian formations of the Hungarian Danube Basin (Fig. 4.5) are traditionally subdivided into low-grade (Szentgotthárd Phyllite, Mihályi Phyllite, Bük Dolomite, Ölbö Carbonatephyllite) and very low-grade formations (Nemesklosta Sandstoneschist, Sótony Metabasalt). All of them belong to the Rábamente Metamorphics. On the northwestern flank of the Transdanubian Central Range, the Vaszar Metamorphics succession includes the Tét Shale and the Vaszar Metabasalt formations, which in some classifications (e.g. Füllöp, 1990) are considered to be part of the Balaton Phyllites (see Fig. 4.6). In contrast to other formations in the study area, all of these formations occur in the subsurface (Fig. 4.3). Correlation attempts with the much better known Paleozoic region in Austria outcropping near Graz and in the basement of the Styrian Basin is discussed in Appendix A.

The Szentgotthárd Phyllite was penetrated in two boreholes (Szentgotthárd(Sz tg)-1, -2) on the eastern flank of the South Burgenland swell close to the Austrian-Hungarian border (Fig. 4.3). The dark grey, psammite, silty slate in Sz tg-2 reflects very low-grade metamorphism, while the sericite-calcite-chlorite schist in Sz tg-1 shows the characteristics of low grade (greenschist) metamorphism (Árkai et al., 1987). The "mixed" K/Ar age (143 Ma, white mica) obtained by Árkai and Balogh (1989) suggests moderate
Fig. 4.5. Upper Austroalpine formations of NW Pannonian Basin, beneath the central part of the Hungarian Danube Basin. Based on various sources listed in the text.
Alpine overprint. The >100 m thick formation represents Silurian(?) rocks. The *Mihályi Phyllite* is encountered in a number of boreholes on the Mihályi high (Fig. 4.3). A similar succession was assigned to this formation to the SW of the Mihályi high, near Vát, Ölbö, Pecöl and Ikervár villages (Fülöp, 1990). The lithology of the formation is very diverse, ranging from phyllites to dolomites, with frequent quartzite and greenschist intercalations in a >200 m thick succession. All of these lithologies of perhaps Silurian(?) (Balázs, 1971, 1975) or Middle Devonian(?) (Teleki et al., 1989) age show a low-grade (greenschist facies) Alpine metamorphism (Árkai et al., 1987).

The *Bük Dolomite, Ölbö Carbonatephyllite, Nemeskóta Sandstoneschist* were drilled in numerous boreholes around Bük, Ölbö, Rábasömjén, Nemeskóta and Ikervár (Figs. 4.3 and 4.5). These lithologies differ from the Mihályi Phyllite by having a very low, metamorphic grade (Árkai et al., 1987), which is also reflected by the lesser reset (between 203 and 140 Ma) or even by a Hercynian K/Ar age (314 Ma, Árkai and Balogh, 1989). In spite of this marked difference, Fülöp (1990) considered these formations as being stratigraphically below the Mihályi Phyllite. In any case, the primary age of these formations spans probably the Ordovician-Silurian (Balázs, 1975) or the Silurian-Devonian interval (Teleki et al., 1989). On the SE flank of the Ikervár high in the Sótony-2 borehole, a slightly altered basalt sequence was drilled and named *Sótony Metabasalt* by Balázs (1971).

The *Tét (or Vaszar) Shale* in the NW flank of the Bakony Mts. near Takácsi and Vaszar (Figs. 4.3 and 4.6) represents a succession comparable to those found in the area of Nemeskóta and Ikervár. In the Tét-2 borehole very low-grade rocks were found below a pronounced erosional unconformity separating the overlying nonmetamorphic Permo-Mesozoic section from the pre-Alpine basement (Balázs, 1971). Indeed, Hercynian K/Ar ages from the Tét Shale cluster around 320 Ma (Árkai and Balogh, 1989). The *Vaszar Metabasalt* has the same lithology as the Sótony Metabasalt (cf. Figs. 4.5 and 4.6).
### Fig. 4.6. Paleozoic formations of the Upper Austroalpine of the NW Pannonian Basin in Hungary. Based on various sources listed in the text.

<table>
<thead>
<tr>
<th>Time in Ma (T.)</th>
<th>Chrono-Stratigraphy</th>
<th>Paleozoic Lithostratigraphy</th>
</tr>
</thead>
<tbody>
<tr>
<td>250</td>
<td>SCYTHIAN</td>
<td>LITTLE HUNGARIAN PLAIN</td>
</tr>
<tr>
<td></td>
<td>ZECHSTEIN</td>
<td>BALATON HIGHLAND</td>
</tr>
<tr>
<td></td>
<td>ROTLIEGENDES</td>
<td>VELENCE MTS.</td>
</tr>
<tr>
<td>-300</td>
<td>GZELIAN</td>
<td>NADASKUT DOLOMITE 20-50</td>
</tr>
<tr>
<td></td>
<td>KASIMOVIAN</td>
<td>DINNYES DOLOMITE 100-200</td>
</tr>
<tr>
<td></td>
<td>MOSCOVIAN</td>
<td>BALATONFELVIDEK SANDSTONE</td>
</tr>
<tr>
<td></td>
<td>BASHKIRIAN</td>
<td>TABAJD EVAPORITE 400-500</td>
</tr>
<tr>
<td></td>
<td>SERPUKHOVIAN</td>
<td>KEKKUT DACITE 600</td>
</tr>
<tr>
<td></td>
<td>VISEAN</td>
<td>GEISE QUARTZDORITE</td>
</tr>
<tr>
<td></td>
<td>TOURNASIAN</td>
<td>DINNYES GRANODIORITE</td>
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<tr>
<td></td>
<td>UPPER</td>
<td>VELENCE GRANITE</td>
</tr>
<tr>
<td></td>
<td>MIDDLE</td>
<td>FÜLE CONGLOMERATE 600</td>
</tr>
<tr>
<td></td>
<td>LOWER</td>
<td>SZEKESFEHÉRVÁR LIMESTONE</td>
</tr>
<tr>
<td></td>
<td>PRIDOLI</td>
<td>100-200</td>
</tr>
<tr>
<td>-450</td>
<td>LUDLOW</td>
<td>REVFULÓP SLATE 200-500</td>
</tr>
<tr>
<td></td>
<td>WENLOCK</td>
<td>SZEKESFEHÉRVÁR LIMESTONE</td>
</tr>
<tr>
<td></td>
<td>LLANDOVERY</td>
<td>ÚRVIDA LIMESTONE 220</td>
</tr>
<tr>
<td></td>
<td>ASHGILL</td>
<td>POLGÁRD MARBLE 200</td>
</tr>
<tr>
<td></td>
<td>CARADOC</td>
<td>REVFULÓP METAANDESITE</td>
</tr>
<tr>
<td></td>
<td>LLANDEILO</td>
<td>UDÁG MARBLE 200</td>
</tr>
<tr>
<td></td>
<td>LLANVIRN</td>
<td>TESEMARLIMESTONE 200</td>
</tr>
<tr>
<td></td>
<td>ARENIG</td>
<td>BALATONFÖKÁJÁR QUARTZPHYLLITE 200</td>
</tr>
<tr>
<td></td>
<td></td>
<td>KOSZÁRHÉGY SLATE 800</td>
</tr>
</tbody>
</table>
The anchimetamorphic rocks drilled in the NW flank of the Bakony Mts. are comparable to those which actually outcrop just to the N of Lake Balaton, in the Balaton Highland (Fig. 4.3). Their correlation was suggested by a number of workers (Oravecz, 1961; Lelkes-Felvári, 1978; Árkai and Lelkes-Felvári, 1987).

The pre-Permian Paleozoic formations of the southeastern flank of the Transdanubian Central Range can be subdivided into several major formation groups (Fig. 4.6). The distribution of the Lower Ordovician - Lower Devonian Balaton Phyllite Formation group is restricted to the Balaton Highland and the Velence Mts. A number of Devonian carbonates are also distinguished in the area. Various formations of Carboniferous age were described from the area between Lake Balaton and Lake Velence, and granitoid intrusions are aligned parallel with the so-called Balaton shear zone (Fig. 4.3). Finally, the lower Permian is represented by a volcanic formation, which was found in boreholes in the Bakony Mts.

The Köszárhegy (or Szárhegy) Slate is known only from boreholes, and it was drilled only in the area of Szabadbattyán (Fig. 4.3). This formation represents the oldest rocks in the study area, which are Lower Ordovician (upper Arenigian) in age (Albani et al., 1985). Its lithology is dominated by very low-grade slates, metamorphosed during the Hercynian orogeny. The formation has a >875 m thickness, and underlying formations were never reached (Lelkes-Felvári, 1978). The Upper Ordovician Balatonfőkajár (or Főkajár) Quartzphyllite has only one single outcrop near Balatonfőkajár, but it was found in numerous boreholes in the area between Lake Balaton and Lake Velence. Its lithology is dominated by alternating quartzphyllite and sericitephyllite. The typically 200 m thick formation has a low-grade, greenschist facies metamorphic overprint of Hercynian age. Locally, e.g. at Balatonhídweg (Fig. 4.3) high grade metamorphism occurred in almandine-amphibolite facies (Árkai, 1987), due to contact metamorphism near the Hercynian granitoids of the region (see later).
The Upper Ordovician - Silurian *Lovas Schist* outcrops along the northern shore of Balaton Lake and in the Velence Lake region. The formation is also known from a number of boreholes. Very low-grade (prehnite-pumpellyite-quartz facies) metamorphosed psammitic and pelitic beds (Lelkes-Felvári, 1978) alternate in the typically 400-600 m thick formation. The upper boundary is always an erosional surface, indicating that a significant portion of the formation might be missing. However, Dudko and Lelkes-Felvári (1992) considered an uninterrupted transition to the Devonian Révfüllöp Slate (see below). Black lydite lenses of the *Velence (or Kányásvölgy) Lydite* (Árkai and Lelkes-Felvári, 1987) are less dominant in the lithology. The distribution of the Upper Ordovician *Alsóörs Metarkyolite* is closely related to that of the Lovas Schist (Lelkes-Felvári and Sassi, 1981). The very low-grade (upper part of the anchizone) tuffs, tuffite layers, lavas and subvolcanic bodies (with a thickness on the order of some meters) of rhyolitic and dacitic composition locally show an ignimbritic character (Lelkes-Felvári, 1978). In contrast, the *Litér Metabasalt* outcropping near Litér, in the Balaton Highland, indicates submarine mafic magmatism of Silurian(?) age (Lelkes-Felvári, 1978). The very low to low-grade (upper anchizone-lower epizone) metamorphosed formation is made up of weathered basalt-tuff, basalt or gabbro, chlorite schist with locally subvolcanic, ophitic texture.

The Devonian *Révfüllöp Shale* outcrops along the northern shore of Lake Balaton near Révfüllöp and Kövágóörs (Fig. 4.3). It is composed of psammitic schists, carbonat-sericite schists with autigene breccia levels, volcanic and sandstone intercalations. In contrast to the Lovas Shale formation in this unit, the carbonate content is higher (4-10% on average) and the composition of the volcanic material is distinctly intermediate. The very low-grade (anchizone) formation has an average thickness of 100-200 m up to 550 m (Lelkes-Felvári, 1978). The distribution of the *Révfüllöp Metaandesite* is closely related to the Révfüllöp Shale formation. The geochemistry of the strongly altered, hydrothermally weathered andesite and its tuff indicates a submarine intermediate magmatism.
The Kékkút Limestone was found only in the Kékkút(K)-4 borehole, between 903-1001.1 m, in the Balaton Highland (Fig. 4.3). It is a very low-grade, dominantly sericite-chlorite schist, slate, sandstoneschist. The most peculiar lithology in this formation, however, is the syngenetically folded, red or grey colored limestone and marl, with typical Tentaculites and Conodonta fauna yielding a Lower Devonian (emsian) age (Lelkes-Felvári et al., 1982). The Úrhida Limestone was encountered in the area between Lake Balaton and Lake Velence, in the Úrhida(Ú)-4 borehole (Fig. 4.3). The white, pelagic limestone of Devonian (Emsian, Eifelian and Givetian) age has carbonate sand and autigene breccia intercalations in it. These latter levels show grading and are considered to be calciturbidites. Because of the steep dip (35-70°) in the Úrhida-4 borehole, the real thickness is estimated as about 220 m (Fülöp, 1990). The Devonian Polgárdi Marble outcrops near Polgárdi and Szabadbattyaán (Fig. 4.3). In the course of mineral exploration, the formation was penetrated in a number of boreholes in the same area. The 200 m thick white, thick-bedded crystalline limestone, originally a shallow-marine, peritidal(?) carbonate, shows a low-grade, Hercynian metamorphism (Lelkes-Felvári, 1978).

The Szabadbattyaán Limestone is a Lower Carboniferous (upper Visean) subsurface unit, near Szabadbattyaán (Fig. 4.3). The dominant lithology is a low-grade metamorphosed siltstone, with sandstoneschist and bituminic, graphitic carbonate intercalations. The primary lithology corresponds to shallow marine shales with patch reefs (Lelkes-Felvári, 1978). The lower and upper boundaries of the 100-200 m thick formation are always tectonic, usually a thrust plane (Földváry, 1952). The non-metamorphosed Upper Carboniferous Füle Conglomerate is outcropping near Füle Village (Fig. 4.3). Its dominant lithologies are continental fanglomerate, conglomerate, sandstone and siltstone (Barabás-Stuhl, 1975). The lithology of the pebbles in the polymict conglomerate is quartzite derived from the Balatonfökökajár Quartzphyllite formation (see above). The apparent thickness of the formation was 600 m in the Polgárdi(Po)-2
borehole. The formation is a post-Hercynian, intramontain "molasse".

The Velence Granite, Dinnyés Granodiorite and Gelse Quartzdiorite are Uppermost Carboniferous and partly (?) Lower Permian formations. They outcrop to the N of Lake Velence, in the Velence Mts. (Fig. 4.3). This batholith covers an area of about 15x7 km (Jantsky, 1957). The granitods are known from a number of boreholes to the SW of the Velence Mts. The formation was also encountered to the S of Lake Balaton, in the area of Ságvár (Sá-1,-3). Granodiorite was found to the S of the Lake Velence in boreholes Dinnyés(Di)-3 and Gárdony(Gá)-1a. Granodiorite was also drilled to the S of Lake Balaton, in the Buzsák(Bu-É-1) well (Fig. 4.3). Two dominant lithologies are distinguished: a monzogranite and a porphyritic, biotite-rich granite. The granite has an intermediate potassium content and is considered to be a calcalkaline, S-type, "postkinematic" granite (Buda, 1981). Pegmatite, turmalinegranite and aplite lenses and dykes are frequent.

The non-metamorphic Lower Permian Kékkút Dacite was encountered in numerous boreholes in the Balaton Highland (Fig. 4.3). It is a grey colored, hydrothermally altered lava rock. The chemical composition suggests a dacite, with a slightly high potassium content. About 630 m of tuff and ignimbrite varieties are also known from the Kékkút(K)-4 borehole. The formation is interpreted to be the product of continental "subsequent" magmatism (Fazekas et al., 1981).

4.2 EARLY-ALPINE STRATIGRAPHY (PERMIAN - LOWER CRETAEOUS)

This time interval is represented in all the units of the Austroalpine nappe system and the underlying Penninic units as well (Fig. 4.2).

4.2.1 PENNINIC UNITS

In this chapter the term Penninic refers only to the South Penninic realm; discussion
of North Penninic units such as the Rhenodanubian Zone of the Eastern Alps is excluded. Thus the distribution of the South Penninic successions is restricted to the Rechnitz window group (e.g. Schmidt, 1951) of the Eastern Alps and the Borinka unit (e.g. Plasienka, 1989) in the Little Carpathians (Fig. 4.1).

In the Hungarian part of the Danube Basin, Penninic units outcrop in the eastern part of the Köszeg-Rohonc (Rechnitz) Window (see below) and the Vashegy (Eisenberg) area to the S (Figs. 4.1 and 4.3). From bottom to top four formations are distinguished in this greenschist facies metamorphosed succession (Fig. 4.4). The present-day lithologic column may represent an overturned sequence (cf. Balla, 1984b).

There are metasandstone and quartzite intercalations in the Köszeg Quartzphyllite (Pahr, 1980). The overlying Velem Calcyphyllite formation comprises limestone beds and phyllites. The Cák Conglomerate occurs as intercalations in the upper part of the Velem Calcyphyllite (Oravec, 1979). The conglomerate includes Triassic pebbles and was interpreted as an olistostrome in a deep sea environment (Kázmér, 1986). The Felsőcsatár Greenschist comprises mafic pyroclastics and lava rocks of an ophiolite suite (Koller and Pahr, 1980). The age of greenschist metamorphism is Middle Miocene based on K/Ar (Kubovics, 1983) and fission track dating (Dunkl, 1990, 1992; Dunkl and Demény, in press). An earlier phase of blue-schist metamorphism (Eoalpine?) was also recorded (Lelkes-Felvári, 1982; Koller, 1985).

Some authors (e.g. Balla et al., 1990) consider also the Mihályi Phyllite of the basement of the Hungarian Danube Basin (see above) to be a Penninic succession.

4.2.2 LOWER AND MIDDLE AUSTROALPINE UNITS

The Lower Mesozoic cover of the Lower Austroalpine is very widespread at the eastern end of the Eastern Alps in the Semmering and Wechsel regions (Fig. 4.1). The Tatricum in the Western Carpathians has also a well-developed Mesozoic cover.
The Stangalm Mesozoic just to the W of the Gurktal Complex of the Eastern Alps (Fig. 4.1) represents the thin, tectonically reduced Mesozoic cover of the Middle Austroalpine unit. In the Western Carpathians, the comparable Krížna nappe system of the Middle Austroalpine Fatricum has a more complete Mesozoic sequence.

There is only one indication of the Lower Austroalpine in the basement of the Danube Basin in Hungary. In the Mosonszolnok-1 well (Fig. 4.3) red sandstone was drilled between 2864-2914 m which was considered comparable to the Upper Permian or Lower Triassic of the Tatricum in the Little Carpathians (Körőssy, 1987).

4.2.3 UPPER AUSTRALPINE UNITS

The early Alpine sedimentary succession dominates the Northern Calcareous Alps, but the Upper Austroalpine unit has only small Mesozoic cover sequences in the Gurktal area (Krappfeld) and near Klagenfurt (Viktring). The Mesozoic mass of the Transdanubian Central Range extends below the subsurface to the W under the eastern part of the Danube and Zala Basins (Figs 4.1 and 4.3). To the NE, the Mesozoic Hronic nappes of the Western Carpathians (Choc, Lunz, etc.) are evenly distributed over the deeper tectonic units.

The Upper Permian - Triassic stratigraphy of the Transdanubian Central Range is summarized in Fig. 4.7. The Upper Permian Balatonfelvidék Sandstone outcrops in a more or less continuous belt along the northern shoreline of Lake Balaton. This nonmetamorphosed formation is also known from a number of boreholes drilled for uranium exploration. It was found in wells on the NW flank of the Bakony Mts, in the Tét-2,-5, Csót-1 and Alsószalmavár(Asz)-1 boreholes (Fig. 4.3). Its lithology comprises a few meters to hundred meters thick fanglomerates which can be locally found at the base of the Upper Permian succession overlying the Hercynian unconformity. The fanglomerate is overlain by fluvial conglomerates with a thickness up to 200 m. The dominant
Fig. 4.7. Upper Permian and Triassic formations of the Upper Austroalpine of the NW Pannonian Basin in Hungary. Based on various sources listed in the text.
lithology of the Balatonfelvidék Sandstone is a 300-500 m thick red sandstone and siltstone member with intraformational, resedimented conglomerate bodies, thin (5-20 cm) dolomite and/or gypsum layers, and dolomite concretions. The sedimentary facies of the formation indicates a continental alluvial, fluvial, locally sabkha depositional environment (Majoros, 1983; Majoros in Fülöp, 1990).

The Upper Permian Tabajd Evaporite is a subsurface formation near Lake Velence and the Vértes Mts. (Fig. 4.3). Wells penetrating this formation are: Tabajd(T)-5; Alcsútdoboz(Ad)-2; Dinnyés(Di)-3; Csőr-7 and Gárdony(Gá)-1a. Its lithology comprises siltstone, dolomite, anhydrite and gypsum in several lithologic associations. The thickness of the anhydrite may reach several meters (Csőr-7 well). The 300 m thick formation was deposited in a sabkha environment (Majoros, 1983). The Upper Permian Dinnyés Dolomite is also a subsurface unit in the same area. The 200 m thick formation is a dolomite with oolitic and calcarenitic texture, suggesting an agitated lagoonal paleoenvironment (Majoros, 1983).

The Lower Triassic (Scythian) Arács Marl outcrops in the Balaton Highland. At the NW flank of the Bakony Mts., the formation was encountered in the Csőt-1, Alsószalmavár(Asz)-1, Bakonyszűcs(Bsz)-3 and the Pápateszér-3 borehole. At the SE flank the Kádárta-1, Litér-2, Sóly-7, Iszkaszentgyörgy-3, -4 and the Csőr-7 wells reached the Arács Marl. This lithology includes sandstone, siltstone, shale intercalations and evaporitic dolomites. The 20-50 m thick, shallow marine, lagoonal formation overlies the Permian succession with an unconformity (Balogh, 1981). The distribution of the Lower Triassic (Scythian) Csupak Marl closely follows that of the Arács Marl. It is a siltstone with bioclastic carbonate intercalations. The 50-100 m thick neritic formation is very similar lithologically to the underlying Arács Marl, making the separation difficult. Contact with the overlying Aszőfő Dolomite, however, is very sharp (Bence et al., 1990).

The Lower-Middle Triassic (Scythian-Anisian) Aszőfő Dolomite, aside from its
outcrops in the Balaton Highland, can also be found in the NW flank of the Bakony Mts., as a subsurface unit (e.g. Bakonyszücs-3 well). Macrofaunas are missing from the finely bedded 100 m thick dolomite and evaporitic dolomite, suggesting a hypersaline lagoonal paleoenvironment. The formation has a transitional boundary towards the overlying Iszka Limestone. The Middle Triassic (Anisian) Iszkahegy Limestone has numerous outcrops in the SE Bakony Mts. and the Balaton Highland. In the NW Bakony Mts., the formation was encountered in the wells Alsószalmavár(Asz)-1 and Bakonyszücs(Bsz)-3. A shaly, finely bedded limestone is the dominant lithology of the formation. Less frequent is a nodular variety or one with dolomite lenses in it. Abundant trace fossils are very characteristic for this 30-50 m thick formation (Bence et al., 1990).

The Middle Triassic (Anisian) 100-200 m thick Megyehegy Dolomite forms widespread outcrops in the SE Bakony Mts. It is also a subsurface unit in the NW Bakony Mts. (Bakonyszücs(Bsz)-3 well). Its lithology is very uniform: light grey, thick-bedded carbonate platform dolomite, with autogenic breccia intercalations at its base (Balogh, 1981). In the uppermost part, layers of oncooids were found. The Middle Triassic (Anisian) Felsőrs Limestone commonly outcrops in the SE Bakony Mts. and the Balaton Highland. Its lithology is quite variable. Typical lithologies are dark carbonate marl, marl and dolomitic marl, with nodular limestone intercalations, shaly tuffite, tuffitic, cherty, nodular limestone with frequent ammonite fragments, well bedded limestone with dark grey tuffitic clay. In the deep water (bathyal?), basinal limestone (20 m thick) brecciated, intraclastic and biomicritic textures are the most common (Balogh, 1981).

The name of the Middle Triassic (Ladinian) Buchenstein Formation refers to the classic Alpine section in Tirol (also known as Livinallongo). The 10-200 m thick volcanosedimentary succession is made up of thin-bedded, cherty limestones and numerous tuffitic (pietra verde) intercalations. These latter beds indicate a trachytic-rhyolitic volcanism during the Ladinian (Szabó and Ravasz, 1970; Ravasz, 1973; Cros and
Szabó, 1984). Rocks derived from lava flows and ignimbrites can also be found locally in the Transdanubian Central Range (Horváth and Tari, 1987). The facies of the overlying Ladinian-Lower Carnian Nemésvámos and Füred Limestones indicates deposition in pelagic water depths. The heterotrophic facies of these 20-50 m thick basinal formations is the Diplopora-bearing Budaörs Dolomite, which is a thick (>1000 m) platform carbonate in the NE part of the Transdanubian Central Range (Balogh, 1981).

The 500-800 m thick Carnian Veszprém Marl overlies the Ladinian basinal limestones in the southern Bakony Mts. The formation has a >1150 m apparent thickness in the NW flank of the Bakony Mts. (Bakonyaszics(Bsz)-1 well). Lithologically, it is composed of fine-grained deep water terrigenous rocks, marls, shales with sandstone and limestone interbeddings. Around Veszprém and in the Keszthely Mts. (Fig. 4.3) there are shallow water carbonate intercalations (Raibl Dolomite and Ederics Formation, resp.) in the formation. In the Buda Mts. their heterotrophic facies are the thin-bedded, cherty limestones and dolomites of the Mátyáshegy Formation.

The Upper Carnian to Norian Hauptdolomit Formation is a very thick (1000-1500 m) platform carbonate on top of the Carnian deep water sequence. The lower boundary of the Hauptdolomit is time-transgressive; in the Buda Mts. it is Upper Carnian, while in the Bakony Mts. it is younger, Lower Norian in age (Oravec, 1963). This points to a progradation of the platform to the SW in present-day coordinates. The formation is lithologically homogeneous, aside from local dolomitic Lofer cycles, indicating peritidal origin of the carbonates (Haas, 1985b).

The Norian Kössen Formation is a 50-100 m thick marly unit on top of the platform carbonates of the Hauptdolomit. The lagoonal beds showing pronounced terrigenous influx occur only in the SW Bakony Mts. and their subsurface extension to the Zala Basin (Fig. 4.3). Towards the NE the formation is missing and a platform carbonate directly overlies the Hauptdolomit.
The Uppermost Triassic Dachstein Limestone shows a diachronous relation similar to that of the Hauptdolomit; it began in the Norian in the Buda Mts. and in the Rhaetian in the Bakony Mts. The 200-800 m thick formation is a typical platform carbonate with pronounced Lofer cycles (Haas, 1985). The deposition of the platform carbonate includes the Triassic/Jurassic boundary and the lower boundary of the Hettangian Kardosrét Limestone is drawn at a minor facies change. This change involves the disappearance of the peritidal cycles and the appearance of ooids in the 50-150 m thick limestone. Note that in the NE part of the Transdanubian Central Range there is a small hiatus at the Triassic/Jurassic boundary. The Jurassic stratigraphy of the Transdanubian Central Range is summarized in Fig. 4.8.

The Upper Hettangian to Lower Sinemurian Pisznicce Limestone is a 2-10 m thick bioclastic limestone. It is overlain by the 40-120 m thick, cherty Isztimér Limestone of Sinemurian age. The Pliensbachian Hierlatz Limestone is a crinoidal biocalcarenite. The Pliensbachian Túzkövesárok Limestone is in Ammonitico Rosso facies. The worldwide anoxic event of the Toarcian is represented by the Úrkút Manganese Ore in the Bakony Mts., which forms exploitable deposits near Úrkút and Eplény. The 2-10 m thick Toarcian Kisgercse Marl is not present everywhere. The very condensed, pelagic sedimentation continues into the Aalenian-Bajocian with the Eplény Limestone (30-80 m thick) and the Tőlyhát Limestone (5-30 m). The Bathonian-Callovian Lókút Radiolarite is a massive radiolarite indicating sedimentation below the CCD.

The Oxfordian-Kimmeridgian Pálihálási Limestone is again a 10-15 m thick Ammonitico Rosso. It is overlain by the thinly-beded Szentivánhegy Limestone which developed locally in Hierlatz facies. The Jurassic-Cretaceous boundary can be drawn within the latter formation. The Tithonian-Barremian Mogyorósdomb Limestone is in Biancone facies (i.e. white pelagic limestone with chert) and its partly heterotropic facies
Fig. 4.8. Jurassic formations of the Upper Austroalpine of the NW Pannonian Basin in Hungary. Based on various sources listed in the text.
is the Valanginian-Barremian Borzavár Limestone, a 1-10 m thick biocalcarenite. The stratigraphy of Cretaceous formations is shown in Fig. 4.9.

The Barremian-Lower Aptian Sümeg Marl has a very variable thickness (10-300 m) and its distribution is mostly limited to the SW Bakony Mts. In the NE Bakony Mts., the formation is either present in a reduced thickness or missing. The Aptian Tata Limestone typically unconformably overlies the Triassic to Neocomian succession. This formation is a biocalcarenite (Lelkes, 1985) of variable thickness (2-200 m).

The Neocomian developed in a strikingly different succession in the NE part of the Transdanubian Central Range (Figs. 4.1 and 4.9). In the Gerecse Mts., the Jurassic is overlain by the Berriasian Felsővadacs Breccia composed of calcareous sandstones and breccias. The Valanginian to Lower Hauterivian Bersek Marl with its grey and red marls shows the gradual deepening to bathyal water depth culminating with the deposition of the Upper Hauterivian and Lower Barremian Lábatal Sandstone. This sandstone is interbedded with marls and has a flysch character (Császár and Haas, 1984; Kázmér, 1986). The flysch sequence is overlain by the Upper Aptian to Lower Albian (?) Köszörüköbánya Conglomerate (Sztanó and Báldi-Beke, in press). This turbiditic conglomerate unit was deposited in a deep sea fan environment (Kázmér, 1988; Sztanó, 1990).

4.3 EOALPINE STRATIGRAPHY (MIDDLE CRETAEOUS - PALEOCENE)

This time interval is represented only in the Austroalpine nappe system. The lower boundary of this period is drawn at the beginning of the Albian. The upper boundary is poorly defined and is regarded as the Paleocene/Eocene boundary.

In the Transdanubian Central Range the Albian Alsóperé Bauxite occurs at the base of the Middle Cretaceous sedimentary cycle (Fig. 4.9). 1-5 m thick bauxite lenses occur in the central part of the Bakony Mts. The overlying Middle Albian Tés Claymarl has a
Fig. 4.9. Cretaceous formations of the Upper Austroalpine of the NW Pannonian Basin in Hungary. Based on various sources listed in the text.
much wider distribution along the first-order syncline of the Transdanubian Central Range. The 40-200 m thick formation is made up of fresh-water and brackish-water claymarls (Császár, 1985). To the NE, in the Gerecse Mts. the heterotropic reefal Környe Limestone can be found. The Albian Zirc Limestone is 50 m thick in the N Bakony Mts. and 200 m in the S Bakony Mts. The formation has a number of lithologies (Császár, 1985), but typically it is a reef limestone. The Zirc Limestone has a gradual transition to the overlying unit in the Bakony Mts., but it is erosionaly truncated to the NE in the Vértes Mts. The uppermost formation in the Middle Cretaceous cycle of the Transdanubian Central Range is the Upper Albian to Lower Cenomanian Pénzeskút Marl. The 470 m thick formation has a nodular, glauconitic limestone unit at the base and is overlain by dolomitic marls and by neritic sandstones on top, suggesting a regression (Császár, 1985). The area of the whole Transdanubian Central Range was subaerially exposed during the Upper Cenomanian to Turonian period.

Senonian sediments (Fig. 4.9) are distributed in numerous outcrops on the NW flank of the Bakony Mts. These formations are also known from wells in the same area and farther to the SW, in the subsurface of the Zala Basin (Fig. 4.1). Their deposition began with the Coniacian(?)-Santonian Halimba Bauxite ( Mindszenty, 1984). Although there is no direct evidence for the age of the bauxite, it was presumably deposited just before the Santonian transgression. The karst-bauxite is exploited in several mines ( Mindszenty et al., 1988). The thickness of the overlying Santonian Csehánhya Formation can reach 200 m (Jocha-Edelényi, 1988). Its lithology shows typical fluvial cycles (5-20 m) with conglomerates, sandstones and clays. Although the Santonian Ajka Coal is locally found below the Csehánhya Formation, in the area of Ajka it is the overlying formation (Haas et al., 1985) and consists of four 30 m thick exploitable coal seams.

The 20-100 m thick Campanian Jákó Marl is composed of marls with frequent lumachella interbeds. The basal part of the formation was deposited in a lagoonal
environment related to the ongoing transgression and continued in neritic facies in the upper part. The Campanian-Maastrichtian *Ugod Limestone* is a 100-300 m thick reef limestone, which is partly heterotrophic with the Jákó Marl (Haas, 1979). The limestone is locally made up of rudist remnants. A strong correlation was found between the NE trending Ugod Limestone reefs and the paleomorphology of the pre-Senonian basement (Haas and Jocha-Edélényi, 1979). The overlying Maastrichtian *Polány Marl* was deposited in significantly deeper pelagic water. This locally 500 m thick formation corresponds to the peak of the Senonian transgression. In the topmost part of the Senonian sequence, sandstone intercalations (distal turbidites) occur in the Polány Marl (Haas, 1984).

4.4 MESO-ALPINE STRATIGRAPHY (EOCENE - LOWER MIOCENE)

Since the rocks deposited during this time interval postdate the main formation of the Alpine nappe system, the allocation to Austroalpine units would be meaningless. Thus the description below follows geographic units.

Paleogene sedimentation in the Transdanubian Central Range (Fig. 4.10) began at the base of the Lutetian (Báldi-Beke, 1984). During this time, the sea invaded the southwestern part of the Bakony Mts., forming the 10-30 m thick *Darvastó Formation* (Kecskeméti and Vörös, 1975) as a basal transgressional unit. Its coarse terrigenous clastics (clays, conglomerates) generally overlie Upper Triassic to Upper Cretaceous rocks and locally karst-bauxite lenses (*Gánt Bauxite*) of late Paleocene-early Eocene age ( Mindszenty et al., 1988). The Darvastó Formation grades upward to the neritic 100-200 m thick *Szöc Limestone* (Vörös, 1989). By the end of the Lutetian, however, the carbonate banks drowned (Báldi-Beke and Báldi, 1990). After a transitional marl unit characterized by the abundance of glauconite, the bathyal 50-200 m thick *Pádrag Marl* was deposited (Báldi-Beke and Báldi, 1991). In the upper part of this unit turbiditic
Fig. 4.10. Paleogene formations of the Upper Austroalpine of the NW Pannonian Basin in Hungary. Based on various sources listed in the text.
intercalations can be found. The 100-200 m thick *Halima Tuffit* in the SW Bakony Mts. is the heterotropic facies of the Padrag Marl. It is a bathyal marl with frequent tuffitic intercalations. In the Zala Basin volcanics (*Szentmihály Andesite*) were also found in wells, in the area of Bak and Zalaszentmihály (Körössy, 1989). These are Priabonian in age and coeval with another volcanic center in the Velence Mts. (*Nadap Andesite*). In the latter area, andesite dykes of Middle Oligocene age were also found (Darida-Tichy, 1987).

The transgression reached the northeastern part of the Bakony Mts. and the northwestern part of the Buda Mts. in the Bartonian (Dudich and Kopek, 1980). Here, thick (up to 50 m) paralic coal seams of the *Dorog Formation* were deposited as the basal unit (Vörös, 1989). The coal seams are overlain by a 50-80 m thick neritic limestone unit (*Szépvölgy Limestone*), similar to the Szőc Limestone. In the northwestern part of the Buda Mts. the Szépvölgy Limestone is overlain by a bathyal marl (*Piszke Marl*).

In the Late Priabonian the transgression proceeded to the E. In the northeastern end of the Transdanubian Central Range the typical Priabonian succession shows a basal conglomerate, then a neritic limestone (*Szépvölgy Formation*) grading up to bathyal marl (*Buda Marl*). The Buda Marl is conformably overlain by the Lower Oligocene (Kiscellian) *Tard Clay*. This unit was deposited in an euxinic environment with an upward increasing fresh water influence in its upper part. The sudden appearance of reworked Eocene fossils in the Tard Clay at the base of the NP23 nannoplankton zone (Báldi-Beke, 1977) dates the onset of the extensive "infra-Oligocene denudation" (Telegdi-Roth, 1927) in the uplifted Bakony subbasin, which is responsible for the erosionally truncated sequences in that area. The *Lower Oligocene(?) Iharkút Formation* is an alluvial-fluvial succession, found locally in the Bakony Mts. The 10-50 m thick formation unconformably overlies older rocks including the Eocene Szőc Limestone and is made up of conglomerates and cross-bedded sandstones.

By the end of the Rupelian (Kiscellian), a pronounced, W-directed transgression is
marked by the extensive *Hárshegy Sandstone* sheet (Báldi and Nagymarosy, 1976; Báldi et al., 1976). The Hárshegy Sandstone grades upward into a widespread siltstone, the *Kiscell Clay*, deposited in bathyal water depth (200-1000 m, Báldi, 1986). Turbiditic sandstone intercalations occur quite frequently within the Kiscell Clay, especially in its lower half. In contrast to the underlying Tard Clay, the very thick (up to 1000 m) succession of the Kiscell Clay was deposited during a relatively short time (about 1-2 Ma).

The Rupelian/Chattian (Kiscellian/Egerian) boundary coincides with a basinwide regression, which is correlated (Báldi, 1986; Tari et al., 1992a) with the major mid-Oligocene eustatic sea-level fall of Haq et al. (1987), 30 Ma ago (for a sequence stratigraphic description of the Hungarian Oligocene and Miocene, see Appendix B). In the deeper part of the basin the Kiscell Clay is directly overlain by the *Szécsény Siltstone* deposited still in deep water (i.e. 200-400 m, Báldi, 1986).

At the western basin margin, Chattian siliciclastic formations show a gradual transition from continental to marine deposition (Csatka Gravel, Mány Sand, Törökbálint Sandstone, respectively). In the Bakony Mts., the thick (up to 800 m) fluvial succession of the Csatka Gravel was deposited during Late Oligocene-Early Miocene (~Egerian) (Korpás, 1981). To the E, the Csatka Gravel interfingers with the *Mány Sand* which was deposited in a brackish sedimentary environment (coastal plain). The *Törökbálint Sandstone* is made up of neritic sandstones.

The stratigraphic column of the Miocene and Pliocene in the NW Pannonian Basin is shown in Fig. 4.11. The trend of sedimentation in the Transdanubian Central Range shows an overall regression during the Chattian and Aquitanian (~Egerian). The Törökbálint Sandstone was replaced by the conglomerates and sandstones of the *Budafok Sandstone* at the beginning of the Eggenburgian, displaying a pronounced basinward shift in facies. The continuing regression in the whole Transdanubian Central Range resulted in subaerial exposure by Early Burdigalian (Late Eggenburgian).
Fig. 4.11. Neogene formations of the Upper Austroalpine of the NW Pannonian Basin in Hungary. Based on various sources listed in the text.
4.5 NEO-ALPINE STRATIGRAPHY (MIDDLE MIocene - RECENT)

Very thick sedimentary successions were formed in several basins during the last stage of the Alpine evolution at the junction of the Eastern Alps, the Western Carpathians and the Pannonian Basin. The most important basins are the Inner Alpine, the Styrian, the Zala, the Danube and the Vienna Basins (Fig. 4.1), which can be further subdivided into smaller subbasins (see e.g. Fig. 4.3).

The Neogene formations of the Transdanubian Central Range (Fig. 4.11) lie on earlier rocks with a major unconformity (Bencze et al., 1990). The Bántapuszta Formation consists of Lower to Middle Miocene deposits, separated by an intraformational unconformity marking the Ottnangian/Karpathian boundary. Around Várpalota in the Bakony Mts. (Fig. 4.3), the Bántapuszta Formation is 250 m thick and is composed of sandstones with limestone intercalations.

At the western margin of the Hungarian Danube Basin, in the vicinity of Sopron, the Neogene sedimentary cycle started in Late Burdigalian, above a major unconformity. Here traces of the Gyulakeszi Rhyolitetuff (19.0±1.4 Ma, Hámor et al., 1987) can be found, defining the base of the Ottnangian regional stage. The overlying Upper Ottnangian Brennberg Coal Formation is separated by the so-called Savian unconformity from the overlying Karpatian sediments. The most widespread Karpatian sediment is the alluvial/fluvial Ligeterdő Gravel outcropping in the area of Sopron.

The Neogene of the deep Zala and Danube Basins (Fig. 4.11) is known only from hydrocarbon exploration wells (Körössy, 1988; Szentgyörgyi and Juhász, 1988). In these areas the Ottnangian is missing and the Middle Miocene (Karpathian) sedimentation began with a basal transgressional conglomerate and sandstone unit. The overlying siltstones are correlated with the Karpathian Garáb Siltstone (or informally "Schlier") indicating relatively deep water (100-200 m). In the Transdanubian Central Range the Garáb Siltstone can be found only locally, e.g. in the Várpalota Basin (Fig. 4.3). This siltstone
laterally interfingers with neritic sandstones (*Budafa Formation*) in the deep basins.

The Karpatian/Badenian boundary is placed at the so-called *Styrian* unconformity marked by the *Tar Dacitetuff* (16.4±0.8 Ma, Hámor, 1985). This tuff is a widespread, sometimes reworked, strongly weathered 1-10 m thick tuff layer. The overlying Badenian deep water *Báden Clay* is separated by the so-called *Leitha* unconformity from the *Pusztamiske Formation*. This latter consists of abrasional conglomerates at its base and neritic sandstones higher in the 100-200 m thick formation. It commonly has a green color due to abundant glauconite. The coal seams of the *Hidas Coal* are found in two horizons in the Bakony Mts., near Herend and Várpalota (Fig. 4.3). The deeper water, 100-120 m thick *Szilágy Claymarl* is the partly heterotropic facies of the Pusztamiske Formation. The claymarl laterally interfingers with the *Rákos Limestone* which is a *Lithothamnium*-bearing neritic limestone. This transition was also frequently observed in the subsurface of the Zala Basin.

The 10-50 thick Sarmatian *Kozárd Formation* was deposited in the deeper part of the basin. The heterotropic *Tinnye Formation* comprises a neritic limestone with bentonite intercalations (*Galgavölgy Formation*, 13.7±0.8 Ma, Hámor, 1985). The 100 m thick continental-limnic *Gyulafirátót Formation* with its variegated clays partly overlies the Tinnye Formation and it marks the end of the Middle Miocene sedimentary cycle. In contrast to the Badenian, the Sarmatian coarse limestones, siltstones and sands were formed in brackish water.

The several kilometers thick Upper Miocene to Pliocene (Pannonian, s.l.) lake sediments unconformably overlie the Sarmatian sequence in the basinal areas (*Zala Marl*) while they are transgressive on older rock units at the basin margin (*Mihályi Conglomerate* and *Ősi Variegated Clay*). In the deep basins (Danube and Zala) two thick (locally up to 1000 m) turbidite systems (*Tőfej Sandstone*) were distinguished interbedded in deep basinal marls (*Lenti Marl, Nagylengyel Claymarl*). Higher in the section,
sandstones (Újfalu Sandstone) and marls (Dráva Claymarl) are alternating within a 300-600 m thick interval indicating delta slope, then delta plain sedimentation. The delta complex is overlain by a 100-600 m thick fluvial succession (Hanság and Rábaköz Formations) which is eroded due to post-Pannonian uplift.

In the Transdanubian Central Range the Uppermost Miocene to Pliocene formations (Pannonian s.l., Fig. 4.11) are subdivided into two major groups (Jámbor, 1980, 1989). The Lower Pannonian corresponds to the Peremarton formation group and the Upper Pannonian is represented by the Dunántúl group. The Peremarton formation comprises from bottom to top the Ós Variegated Clay, Kisbér Gravel, Szák Claymarl and Csákvár Claymarl formations. The cumulative thickness of these brackish, clastic formations is on the order of several hundred meters. The Dunántúl formation group is made up of the Kálla Gravel, Somló Formation, Kapolcs Limestone, Taliándörögd Marl, Tihany Formation and Nagyvázsony Limestone formations.

The Tapolca Basalt formation includes the widespread Pliocene (Balogh et al., 1985) basalts and tuffs of more than 60 eruptive centers (Jámbor et al., 1981) scattered over the Bakony Mts. and the Balaton Highland (Fig. 4.1). The basalt has a pronounced alkaline character (Szabó et al., 1992) and is considered to be "post-orogenic".

Above an erosional unconformity the Quaternary sediments are represented by tens of meters thick Pleistocene loess sequences. The bulk of the study area shows recent uplift and erosion, but the center of the Danube Basin between Mosónmagyaróvár and Győr (Fig. 4.3) is still strongly subsiding as shown by an up to 600 m thick Quaternary fluvialite sedimentary sequence.

4.6 PRE-ALPINE FACIES AND STRUCTURAL EVOLUTION

Although the pre-Alpine facies and structural evolution will not be addressed in the next chapters, I have briefly summarized the current knowledge. To date very little is
known about the pre-Alpine evolution of the basement (Fig. 4.3) underlying the thick Neogene basin fill of the Danube Basin (Fig. 4.12). Structural studies in recent years, however, provided a considerable amount of data and geodynamic interpretations for the Eastern Alps and the Western Carpathians (see Appendix A).

The basement of the Hungarian Danube Basin in the area of Nemeskolta and Takácsi consists of Paleozoic rocks (see 4.1.2 subchapter) that yielded Hercynian metamorphic ages (Árkai and Balogh, 1989). The K/Ar ages of the very low-grade rocks were between 311 Ma and 329 Ma and were related to the Sudetian phase.

In the southern flank of the Transdanubian Central Range, in the area between Lakes Balaton and Velence, Dudko (1988) carried out modern microtectonic measurements on the very low-grade metamorphosed Paleozoic Balaton Phyllite group (see 4.1.2 subchapter). She reconstructed a Hercynian stress field with a maximum compressive axis oriented NNE-SSW. During the first phase of deformation, isoclinal folding with pronounced axial-plane schistosity developed, followed by formation of SSW-vergent nappes and associated folds. The unmetamorphosed Upper Carboniferous Füle Conglomerate postdates this deformation.

4.7 EARLY ALPINE FACIES AND STRUCTURAL EVOLUTION

The Permian to Early Cretaceous evolution of the NW Pannonian basin is subdivided into a Triassic transtensional period and a subsequent Jurassic extensional period. They are described below in terms of facies evolution, as very few early Alpine structural features are known.

4.7.1 TRIASSIC "ABORTED" RIFTING

The term "aborted" rifting (Bechstädt et al., 1978) is adopted in this subchapter, since in the NW Pannonian Basin Middle Triassic rifting did not advance to the opening of
Fig. 4.12. Geological section of Körössy (1981) across the study area. For location see Fig. 4.3. Vertical exaggeration is tenfold.
an oceanic basin, in contrast to the Meliata-Bükk area just E of the study area (see Chapter 2).

The best known part of the Transdanubian Central Range is the Bakony Mts. (Fig. 4.1). Here, the Upper Permian terrestrial sandstones were deposited near base level, and the onset of marine sedimentation occurred during earliest Triassic times. As for the whole Alpine realm, the sea of the Paleothethys transgressed the area from the E (e.g. Laubscher and Bernoulli, 1977). The lowermost Triassic beds show a lagoonal-supratidal depositional environment followed by marls. The appearance of marls indicates the gradual widening of the basin but not its deepening, since the marls are overlain by dolomites with evaporitic intercalations that indicate lagoonal sedimentation (Galácz et al., 1985). The Lower Triassic succession is overlain by the Anisian Megyehegy Dolomite. This evenly-thick platform carbonate shows that the whole basin had a uniform, shallow-water character.

A sudden change occurred during the late Anisian, when the platform carbonates were replaced by basinal cherty limestones and marls. These lithologies are similar to the Recoaro and Reiffing limestones of the Southern and Eastern Alps and suggest disintegration of the previous platform by block-faulting (Galácz et al., 1985; Budai and Vörös, 1992). Numerous acidic tuff horizons around the Anisian/Ladinian boundary are associated with nodular limestones and claystones and correspond to the Buchenstein beds of the Alps (Cros and Szabó, 1984; Budai, 1992). This succession represents a Middle Triassic rifting period (Bechstädt et al., 1978; Budai and Vörös, 1993). In the Bakony Mts., aside from widespread tuffs, some lavas occur. The alkaline character of the volcanites (Horváth and Tari, 1987) suggests rifting. The whole geodynamic picture, however, suggests transtensional basin opening in a wide sinistral shear zone.

The late Ladinian red nodular limestone of the Bakony Mts. suggests the stabilization of basinal environment. At the same time, however, very thick platform
carbonates deposited in the NE part of the Transdanubian Central Range survived the Middle Triassic extensional period (Balogh, 1981).

During the Carnian more block-faulting occurred. This is indirectly suggested by the abrupt appearance of marls with calciturbidite intercalations. In the Bakony segment of the Transdanubian Central Range, these marls dominate in the northern part. Allodapic carbonates are more frequent in the southern part, suggesting a platform to the S. Higher in the section, neritic carbonates indicate the progressive infilling of the Carnian basin by sediments derived from surrounding platform areas.

By the late Carnian all the basins were filled by the peritidal Hauptdolomit. This thick dolomite succession is overlain by the marly Kössen Beds at the base of the Rhaetian, suggesting a period of strong terrestrial influx. In the NE part of the Transdanubian Central Range, however, the Hauptdolomit is directly overlain by the platform Dachstein Limestone. The Triassic/Jurassic boundary is drawn within this intertidal limestone as a minor facies change that heralds the subsequent drowning of the Triassic platforms.

4.7.2 JURASSIC OCEANS

The Jurassic ocean of the Tethys opened to the N of the aborted Triassic rifts. The sediments of this deep oceanic basin can be found in different structural units of the study area. The actual oceanic crust, however, is known only from the Rechnitz Window group.

It is also important to note that in contrast to the Eastern Alps and the Western Carpathians, Jurassic facies changes cannot be observed in a N-S or NW-SE direction across the Transdanubian Central Range, because these rocks are preserved only in a narrow NE-trending strip along the axis of the range (Fig. 4.13). Therefore facies belts are distinguished only along a NE-SW line.

By the end of the Hettangian the carbonate platform disintegrated, as suggested by
Fig. 4.13. Tectonic map of the Transdanubian Central Range (Dudko, 1992a). For location see Fig. 4.3.
various heterotrophic carbonate facies (Galácz et al., 1985). The deposition of these facies occurred respectively in basinal, seamount-marginal and seamount environments following extensional block-faulting. Even the tops of seamounts drowned below the photic zones as suggested by the carbonate facies.

The appearance of the first red, pelagic sediments over the platform carbonates is markedly diachronic along the NE-trending Transdanubian Central Range (Kázmér, 1987a). The Zala Basin shows the most pronounced subsidence, suggested by Liassic black marls. During the Liassic, this deep basin was separated by the submarine Bakony Plateau from another relatively deep basin in the Vértes and Gerecse Mts.

Sequences that overlie the drowned platform show a gradual deepening with increasing pelagic character ending with the appearance of the characteristic Ammonitico Rosso. Due to ongoing subsidence the deeper parts of the uneven basement sank below the CCD, resulting in the diachronous replacement of the pelagic limestones by massive radiolarites from the Toarcian onward (Galácz, 1976). During the Oxfordian the Ammonitico Rosso reappeared at the same time in the whole area, suggesting a regional fall of the CCD (Galácz et al., 1985) or perhaps changing current systems.

During the Tithonian the red pelagic limestones were uniformly replaced by the Biancone facies of the SW part of the Transdanubian Central Range. The deposition of these white, cherty pelagic limestones that become more argillaceous and less siliceous upward is characteristic for almost the whole Early Cretaceous (Császár and Haas, 1984). In the Sümeg area, E of the Zala Basin (Fig. 4.3), a thick marl unit also shows the gradual infilling of the basinal areas.

In contrast to the more than 500 m thick Neocomian sequence of the Zala Basin, in the Bakony Mts., a few meters thick section was formed during the same period. This suggests condensed deposition on a submarine plateau (Kázmér, 1988). In the NE part of the Transdanubian Central Range, however, a flysch trough existed at the same time.
(Császár and Haas, 1984). Paleotransport directions in the more than 300 m thick flysch sequence suggests a source area to the present day NE (Kázmér, 1987). Significant amounts of chrom-spinel in this sequence indicates an ophiolitic provenance (Árgyellán, 1989), supposedly from the Vardar suture (Sztanó, 1990).

The Lower Cretaceous of the Transdanubian Central Range ended with the widespread deposition of a neritic crinoidal limestone (Tata Limestone) which locally unconformably overlies the older sequences (Fülöp, 1964). The abundant extraclasts in this rock are locally derived from older Mesozoic rocks (Mészáros, 1971) suggesting that the last remnants of the former uneven topography were eroded during this time. The uplift and erosion are traditionally regarded as the first sign of the Eoalpine compressional period (e.g. Galácz et al., 1985). Thus the deposition of the Tata Limestone dates also the initiation of the first-order syncline of the Transdanubian Central Range (Fig. 4.13), which controlled the subsequent facies patterns during Middle Cretaceous times (Császár, 1985).

4.8 EOALPINE FACIES AND STRUCTURAL EVOLUTION

The Eoalpine evolution is subdivided into two periods. During the Middle Cretaceous large-scale overthrusting and the initial formation of the Austroalpine nappes occurred. Sediments deposited during the Late Cretaceous and Paleocene seal these Eoalpine s.s. nappe contacts. Several markedly different basin-forming mechanisms were proposed for these "post-tectonic" Gosau successions.

4.8.1 MIDDLE CRETACEOUS NAPPE TECTONICS

To date very little is known about Middle Cretaceous compressional structures of the Eoalpine evolution of the Transdanubian Central Range. Such structures are much better known from the Eastern Alps and the Western Carpathians.

Cretaceous compressional structures have been known from the Bakony Mts. and
the Balaton Highland since the last century (e.g. Böckh, 1874; Laczkó, 1911; Lóczy, 1913; Teleki, 1936; Erdélyi-Fazekas, 1943). But there is an ongoing debate on the timing of these structures and perhaps more importantly whether or not they represent an Alpine nappe structure of the Transdanubian Central Range or not. The different views are reviewed below after a brief summary of the facies evolution.

The first Eoalpine compression in the Aptian resulted in the uplift and subareal exposure of the flanks of the first-order syncline of the Bakony Mts (Fig. 4.13). Bauxites (Alsópere Bauxite) developed over the strongly karstified Mesozoic carbonates, during the Albian ( Mindszenty, 1984; Maksimovic et al., 1991). This karstified surface was invaded by the sea from the NE (Császár, 1985).

In the Gerecse Mts., however, there is a continuous transition of Aptian Tata Limestone into the overlying pelagic siltstone (Vértessomló Siltstone). This basinal rock interfingers to the SW with reefal limestones (Környe Limestone) in the Vértes Mts. (Császár and Haas, 1984). In the Bakony Mts., brackish clays and littoral sandstones gradually transgress to the SW. This deepening coincided with the appearance of reefs (Zirc Limestone) over the brackish successions of the Bakony and Vértes Mts. (Császár, 1985). These reefs are overlain by Upper Albian to Cenomanian marls in the Bakony and Vértes Mts. with a small hiatus that increases to the NE (Fig. 4.9, Császár and Haas, 1984). Finally, by the end of the Cenomanian the whole Transdanubian Central Range was subareally exposed and erosion took place until the Santonian when the Senonian sedimentary cycle started.

Opinions differ about the beginning and the main phase of Eoalpine compression in the Bakony Mts. According to Mészáros (1983), the onset of folding was at the Hauterivian/Barremian boundary, because the Barremian locally overlies folded Neocomian-Jurassic-Triassic strata with an unconformity. Mészáros (1971) documented large Jurassic carbonate boulders in the Tata Limestone of the Csehbánya Basin (Fig. 4.3)
and suggested strike-slip activity during the Aptian. For most authors, however, compression began in the Early Albian (e.g. Császár, 1985) corresponding to the Austrian phase of Stille (1924, see later). In Hungary the term Tisian was adopted to denote this phase (e.g. Fülöp, 1964). For other workers the folded sequence includes the Cenomanian, and thus the Eoalpine compression took place from the Late Cenomanian to the Santonian (e.g. Dudko, 1992b).

Note another point about the first-order syncline of the Transdanubian Central Range (Fig. 4.13). On several published maps (e.g. Haas and Jocha-Edelényi, 1980, Dudko, 1992b), one sees two Early Cretaceous synclines in a right-stepping en échelon geometry in the axis of the Bakony Mts. One of them is the Zirc-Úrkút and the other is the Magyarpolány-Sümeg syncline. Furthermore, the map of Dank and Fülöp (1990) indicates a third syncline in the area of Nagytalaj (Fig. 4.12). It is therefore more appropriate to talk about several second-order synclines within the Bakony Mts.

Two major thrust faults are known in the area since the beginning of the century. Both are located in the Balaton Highland (Fig. 4.14); the northern one is called the Veszprém (Laczkó, 1911; Erdélyi-Fazekas, 1943) and the southern one is the Litér thrust (Böckh, 1874; Lóczy, 1913; Teleki, 1936). These thrusts have a SE vergence and can be traced along strike for more than 100 km. Dudko (1991) studied the microstructures associated with these thrusts and suggested the flattening of these faults with depth into hypothetical detachment planes within the Hercynian basement. Dudko (1992) recently proposed about 1.5-2 km thrust displacements on both faults but admitted that the real value may exceed these figures considerably. Dudko (1991) also showed an additional southern and a northern thrust parallel to the previous thrusts (Fig. 4.14). Balla and Dudko (1989) extrapolated the subsurface continuation of the Veszprém and Litér thrusts to the NE up to the Buda Mts. (see Fig. 4.13).

While these thrusts and associated anticlines (e.g. Lóczy, 1913; Budai, 1991) can be
Fig. 4.14. Structure of the Balaton Highland (Dudko, 1991). For location see Fig. 4.3.
mapped on the surface, there are also indications of major thrusts verging to the NW on the NW flank of the Bakony Mts., i.e. in the subsurface of the Danube Basin. Some examples were best shown on reflection seismic sections mentioned in Chapter 3 (e.g. Horváth and Rumpler, 1984; Horváth et al., 1987; Rumpler and Horváth, 1988; Pápa et al., 1990; Mattick et al., 1989). Unfortunately the supposed overthrusts were never penetrated by drilling.

A much debated question is whether the thrusts interpreted on seismic data from the NW flank of the Bakony Mts. and the Litér, Veszprém and other smaller thrusts in the Balaton Highland should be seen as the manifestation of a large Eoalpine nappe structure. Balla's (1992) answer is a definite no while Dudko (1992) suggests that this problem cannot be solved at present. A number of authors following Horváth and Rumpler (1984) and Horváth et al. (1987) accepted the large-scale allochthony of the Bakony Mts. as a working hypothesis (e.g. Fülöp, 1990; Tari, 1991). It is an important objective of this thesis to document this inferred allochthonous structure (see Chapter 5).

Regardless of the uncertainties concerning the significance of the above Eoalpine thrusts and the Early Cretaceous synclines, all are postdated by a major bend in the strike of the Transdanubian Central Range. This bend changed the NE-trending strike of the Bakony and Vértes Mts. to an E-trending one in the Gerecse Mts. and farther to the SE to a NW-trending strike in the Buda Mts (Fig. 4.13). The age of the bend is Middle Cretaceous (e.g. Wein, 1977) and probably predates the intrusions of early Senonian lamprophyre dykes in the area (Balla, 1988). This bending is an important element of the Eoalpine evolution of the Alps if the escape model of Kázmér and Kovács (1985) is correct. The tectonic mechanism that formed this large-scale bend is not known (Balla, 1988). A possible explanation is offered in Chapter 5.

The formation of the lamprophyre dykes (Horváth et al., 1983; Horváth and Ódor, 1984) corresponds to a tectonic phase taking place just before the Upper Cretaceous. This
pre-Gosau phase is regarded as an extensional event in the Bakony Mts., since it also produced numerous extensional dykes in the Aptian limestones that were filled by red calcite, e.g. in the area of Sümeg (Haas et al., 1985). This red calcite can be found in pebbles of the basal transgressive conglomerate of the overlying Senonian strata.

4.8.2 SENONIAN "GOSAU" BASINS

Senonian basins cover a significant portion of the NW Pannonian Basin with the surrounding parts of the Eastern Alps and Western Carpathians. Only part of them, however, can be called Gosau basins s.s., a term which refers to a specific siliciclastic facies developed in the Northern Calcareous Alps. The Senonian basin of Western Hungary has a different, dominantly carbonate facies from that of the Gosau.

A Senonian basin is located to the SE of the Rába Line and to the NW of the axis of the Bakony Mts. (Fig. 4.3). Following the Middle Cretaceous emergence and erosion, during the Santonian the sea invaded the area of the Bakony Mts. from the SW (Haas, 1979). Prior to the transgression over the strongly karstified surface, bauxite lenses were deposited in topographic lows. The bauxite is overlain by terrestrial beds (Jocha-Edelényi, 1988) with intercalated coal seams (Haas et al., 1986). During the continuing subsidence, shales and marls were deposited in lagoonal environments.

Note that deposition of the basal terrestrial to lagoonal beds was controlled by an inherited, pre-Senonian topography (Knauer and Gellai, 1978; Haas and Jocha-Edelényi, 1979; Haas, 1983). The reconstruction of this paleotopography is based on the isopach of the terrestrial-lagoonal succession determined by numerous coal exploration wells in the Ajka area (Fig. 4.15). This reconstruction shows that the area was characterized by some NE-trending paleohighs emerging above the enclosed elongated basins by about 250 m. The average spacing between adjacent highs is about 10 km. Haas (1983) regarded these highs as horsts. Interestingly enough, this topography also controlled the appearance and
Fig. 4.15. Reconstruction of the Early Santonian paleomorphology in the northwestern flank of the TCR (Haas et al., 1986). For location see Fig. 4.3.
spatial distribution of the widespread Campanian rudist reefs (Haas, 1979). On the paleohighs, the reefs lie directly on top of the pre-Senonian basement with lateral transitions to the basinal lithofacies deposited in the paleolows (Haas and Pálfalvi, 1989).

By end Campanian, the reefs drowned due to the gradual deepening of the basin, and pelagic marls were deposited throughout the whole basin (Haas, 1983). In these marls, coarse breccias can be found locally which are interpreted as submarine fan deposits derived from adjacent platform areas. In the uppermost Maastrichtian part of the section thin siltstone and sandstone intercalations indicate increasing terrestrial influx heralding a regression in the basin (Haas, 1983). The post-Maastrichtian evolution of the Senonian basin is not known because of severe erosion during the Paleocene-Early Eocene.

As pointed out by Haas (1985a), the above described Senonian strata differ from the coeval Alpine Gosau successions and should not be regarded as genuine Gosau deposits. The dominance of thick and widespread reefal limestones is in strong contrast to the dominantly clastic and flyschoid Gosau basins of the Eastern Alps (e.g. Faupl et al., 1987). Haas (1983) considered the Hungarian Senonian basin to be an "epicontinental" basin without suggesting any specific basin-forming mechanism.

Recently, Tari (1992) tentatively interpreted the Senonian basin of NW Hungary in terms of a flexural basin (see Chapter 5).

4.9 MESOALPINE FACIES AND STRUCTURAL EVOLUTION

The Mesoalpine period is discussed in two subchapters below. First, the evolution of the Eocene to Early Miocene basins is outlined using a previously proposed model. Second, the Mesoalpine magmatites and their interpretations are discussed.

4.9.1 PALEogene "EPICONtinental" BASINS

The distribution of these basins is similar to the Senonian "epicontinental" basins in
the same area. The Paleogene basin of the Transdanubian Central Range, however, is larger than the Senonian basin and at one time covered the whole range. The Krappfeld Eocene basin in the Eastern Alps (Fig. 2.4) as an isolated erosional remnant coincides with the underlying Senonian basin.

The Middle to Late Eocene evolution of the Paleogene basin in W Hungary is similar to the Senonian basin evolution. At the beginning of the Lutetian, the sea invaded the western part of the Bakony Mts. from the SW. The transgression took place on karstified Mesozoic terrains, where the basal clastics frequently overlie bauxite lenses which were in part deposited during the Late Paleocene-Early Eocene (Mindszenty et al., 1988). The deposition of bauxites was controlled by normal faults. As a result of the ongoing subsidence the clastic sequence (with local coal beds) is overlain by neritic carbonates deposited on a NW-dipping ramp. By the end of the Lutetian, however, the carbonate banks were drowned, first in the northwestern, and then in the southeastern part of the SW-Bakony with an approximate time lag of about 2 Ma (Báldi-Beke and Báldi, 1990). The overlying bathyal marls (Padrag and Piszke Marls) were deposited in a 800-1200 m deep sea suggested by abundant planktonic foraminifera (Horváth-Kollányi and Nagy-Gellai, 1989; Báldi-Beke and Báldi, 1990). Turbiditic intercalations occur in the upper part of these marls. The Bartonian transgression reached the northeastern part of the Transdanubian Central Range in the Buda Mts. (Dudich, 1977; Dudich and Kopek, 1980) and in the Late Priabonian it overstepped to the E of the Danube (Sztrákos, 1975a; Balázs et al., 1981).

Paleogene structures in the Buda Mts. had already been recognized in the first half of the century. Various authors (for a compilation see Balla and Dudko, 1991) reported folds and/or thrusts from the area, which they attributed to the Late Eocene "Pyrenean phase". Horusitzky (1943) postulated Paleogene nappes based on Triassic facies patterns in the Buda Mts. Wein (1977) described Paleogene thrusts and folds from the Buda Mts. These
structures are NE-trending, SE-vergent compressional structures. Balla and Dudko (1991) documented outcrop-scale ENE-striking and SSE-vergent folds in the Lower Oligocene (Kiscellian) Tard Clay from temporary outcrops in Budapest; however, the authors inferred a Miocene age for these compressional folds.

In spite of the limited Paleogene outcrops of the Buda Mts., a large number of field data could be collected including microtectonic and sedimentological observations (Fodor et al., 1992). Several compressional antiforms were mapped which clearly show Late Eocene synsedimentary growth (Fig. 4.16). These NE-trending and SE-verging antiforms are probably underlain by blind thrusts, since no traces of surface-breaking thrusts were found. The blind thrust planes must be connected to a decollement surface at depth, probably within the Triassic succession, suggesting thin-skinned deformation. Microtectonic measurements on numerous outcrop-scale strike-slip faults, reverse faults, tension gashes, etc. revealed a WNW-ESE oriented compressional stress field during Late Eocene-Early Oligocene times (cf. Báldi and Nagymarosy, 1976). Local unconformities due to progressive tilting of the flanks of growing anticlines and syntectonic clastic aprons on the paleoslopes also indicate Late Eocene (Priabonian) tectonic activity in the Buda Mts. (Magyari, 1991).

Báldi and Nagymarosy (1976) defined the NE-trending Buda Line based on abrupt facies changes of Upper Eocene-Lower Oligocene sediments in the Buda Mts. One of the NE-trending Late Eocene anticlines, in fact, may coincide with this "Buda Line", as it appears to separate very different sedimentary facies (Fig. 4.16). This anticline, now called the "Buda Anticline" (Fodor et al., 1992), could be the paleohigh between areas characterized by carbonate deposition in neritic depth and areas of significantly deeper water marl sedimentation to the SE.

The contrasting paleotopography lasted until the Early Oligocene, when to the SE of the Buda Anticline deposition continued with the euxinic Tard Clay, whereas to the NW
Fig. 4.16. Late Eocene structural/sedimentological model of the Buda Mts. showing the progressive steepening of slopes and related (re)sedimentation due to reverse fault propagation in a hypothetical section (Fodor et al., 1992). The Buda Line probably represents a submarine barrier on top of the highest antiform. For an approximate location see Fig. 4.3.
of the Buda Anticline subareal exposure and erosion took place at the same time. The sudden appearance of reworked Eocene fossils in the euxinic clay (Báldi-Beke, 1977) dates this extensive "infra-Oligocene denudation" (Telegdi-Roth, 1927) and the uplifting of the Transdanubian Central Range, which is responsible for the erosionally truncated sequences in that area (Fig. 4.10).

A drastic change occurred in the late Early Oligocene when the sea rapidly invaded the area NW of the Buda "Line". Báldi and Nagy-Gellai (1990) postulated a hypothetical major normal fault responsible for this pronounced transgression marked by an extensive sand sheet. The sandstone grades upward into a widespread siltstone, the Kiscell Clay, deposited in bathyal water depth (200-1000 m, Báldi, 1986).

In the Transdanubian Central Range, there is a definite shift of deposition towards the ENE from the Middle Eocene to the Early Miocene. During the Paleogene a relatively slow subsidence was followed by an abrupt phase of fast subsidence which in turn was followed by uplift, erosion and/or continental deposition (Báldi and Báldi-Beke, 1985). Subsidence and sediment accumulation curves were published documenting this phenomenon (Báldi and Báldi-Beke, 1985; Bernhardt et al., 1988; Vörös, 1989; Báldi and Nagy-Gellai, 1990).

For a long time, the Paleogene succession of the Transdanubian Central Range was described as a molasse basin (e.g. Balázs et al., 1981). Compared to other intra-Carpathian Paleogene basins it was also classified as an epicontinental basin (Nagymarosy, 1990a). In the last decade, however, a pull-apart origin of this basin was proposed by some. Kázmér (1984), Fodor and Kázmér (1989) interpreted the Eocene synsedimentary tectonics in the Buda Mts. in terms of a pull-apart basin. Recently Fodor et al. (1992) extrapolated this model to the whole Transdanubian Central Range. The depocenter migration and the relatively fast deepening of individual subbasins to bathyal water depth was followed by uplift and erosion. The asymmetric and apparently fault-controlled geometry of the basins
led Báldi and Báldi-Beke (1985), Royden and Báldi (1988) to propose also a transtensional pull-apart model.

In contrast to these models Tari (1992) recently interpreted the Paleogene Basin of NW Hungary in terms of a flexural basin (see Chapter 5).

4.9.2 MAGMATISM

Middle to Upper Eocene magmatites are known from the subsurface of the Zala Basin (Körössy, 1988), in the area of Bak and Szentmihály (Fig. 4.3). The locally nearly 1000 m thick succession is made up of alternating tuffs and andesites, suggesting a stratovolcano. NE of Lake Velence, near Nadap (Fig. 4.3), another major Eocene volcanic center can be found (Ďudko et al., 1989). Although sediments intercalated between tuff layers clearly show the Late Eocene age of the volcanism, some andesite dykes in the Carboniferous granite were dated as young as 30 Ma (Darida-Tichy, 1987). Fission track ages on zircons also indicate repeated upper Eocene-lower Oligocene reheating in the Velence Mts. (Dunkl, 1992b). Petrological studies have shown that these predominantly calc-alkaline suites are associated with subduction. More specifically, their geochemistry suggests an intermediate case between an active continental margin type and an oceanic island arc type volcanism (Darida-Tichy, 1987).

The large amount of Eocene andesite pebbles in the Upper Oligocene Csatka Gravel in the Bakony Mts. indicates other volcanic centers hidden beneath the Danube Basin (Korpás, 1981).

In the Buda Mts., Báldi and Nagymarosy (1976) observed very intensive Middle Oligocene hydrothermal activity, (similar to that of the Velence Mts., Darida-Tichy et al., 1983) which is distinctly associated with the Buda Line. This hydrothermal event occurred at the boundary of the NP23/24 nannoplankton zones (~31 or 29.5 Ma, according to the time-scale of Haq et al., 1987 and Harland et al., 1989, respectively).
4.10 NEOALPINE FACIES AND STRUCTURAL EVOLUTION

The Neoalpine evolution of the NW Pannonian Basin is subdivided into three parts. The Middle Miocene syn-rift stage will be discussed separately from the subsequent Late Miocene to Recent post-rift evolution. Finally, some interpretations on the Neogene magmatism related to the opening of the Danube Basins will be discussed.

4.10.1 MIDDLE MIOCENE SYN-RIFT TECTONICS

The three major Neogene basins of the study area are the Danube, the Styrian and the Vienna Basins (Fig. 4.1). The best studied basin is the pull-apart Vienna Basin. To date, the structure of the Styrian and the Danube Basins is much less known.

The debate on the character and significance of the Rába Line will be outlined in this subchapter, even though some regard it as an Oligocene fault. Similarly, the major Middle Miocene strike-slip faults of the Transdanubian Central Range are discussed in this context, although many of them may be reactivated Middle Cretaceous faults.

The Danube Basin of Hungary and Slovakia was first interpreted in terms of a large pull-apart basin by Horváth and Royden (1981). Later, Rumpler and Horváth (1988) also showed this basin to be formed in a divergent strike-slip setting (Fig. 1.9). The pull-apart origin is still supported by some who consider this basin to be analogous to the Vienna Basin (e.g. Tomek and Thon, 1988; Bergerat, 1989). Local Miocene subbasins at the eastern margin of the Hungarian Danube Basin (e.g. Dabrony, Nagygörbő, Fig. 4.3) were also interpreted as pull-apart features (Dudko et al., 1992). As the regional sections A-F in Chapter 3 indicate, many of the subbasins of the Hungarian part of the Danube Basin in fact are bounded by low-angle normal faults rather than strike-slip faults.

A closely related problem is that of the Rába Line trending NE beneath the center of the Hungarian part of the Danube Basin (Figs. 4.3 and 4.12). The Rába Line as such was introduced by Scheffer (1959). In the description of Körössy (1959, 1965) the Rába Line
is a fault of uncertain origin in the pre-Tertiary basement separating a metamorphosed Paleozoic area in the NW (e.g. Mihályi, Ikervár, Takácsi, etc.) from an unmetamorphosed area in the SE (e.g. Dabrony, Vasvár, Borgáta, etc., see Fig. 4.3). Wein (1971) also shared this view. Based on reflection seismic data, Ádám et al. (1984), Horváth and Rumpler (1984), Rumpler and Horváth (1988) interpreted a SE-dipping fault at the southeastern flank of the Mihályi high as the Rába Line separating Paleozoic rocks from probable Triassic rocks in the basement (see Fig. 3.10). Horváth et al. (1987) further interpreted this fault as a flower structure (Fig. 4.17) on the MK-1 line (Fig. 3.9) suggesting oblique-slip movements, i.e. a combination of normal and sinistral strike-slip faulting. This interpretation, however, does not seem tenable as shown on Plate 3, since there are a number of uninterrupted intra-basement reflectors at the site of the supposed flower structure.

Another line of thought is represented by Balázs (1971), Árkai et al. (1987) and Árkai and Balogh (1989). These authors view the Rába Line as a fault separating low-grade metamorphosed Paleozoic rocks (e.g. Mihályi) from very low-grade metamorphosed Paleozoic (e.g. Takácsi, Ikervár) to unmetamorphosed Mesozoic rocks (Fig. 4.3). Mattick et al. (1989) and Teleki et al. (1989) also shared this view.

The difference between the above interpretations is best illustrated on regional line C (Chapter 3.1.3, Plate 1). While Árkai and Balogh (1989) place the Rába Line at about 30 km, to the NW of the Ikervár high, other authors (e.g. Körössy, 1985, Hobot et al., 1990) traditionally put this fault at about 37 km, SE of the Ikervár high.

Further confusion emerged from a different attempt to locate the Rába Line. Magnetotelluric surveys in the Danube Basin revealed a highly conductive intrabasement layer (Takács, 1968) mentioned previously in subchapter 3.1.4. Based on several magnetotelluric transects across the area, the Rába Line was defined as the boundary of the good basement conductor to the NW. The structural feature determined in this way
Fig. 4.17. Line drawing interpretation of the MK-1 line (Horváth et al., 1987). For location see Fig. 4.3.
Asterisks denote the depth of the intra-basement conductor.
(e.g. Pápa et al., 1990), however, does not coincide with the Rába Line of others.

Balla et al. (1990) followed this latter, magnetotelluric approach to argue that the Rába Line is not located at the southeastern flank of the Mihályi high but rather some 8 km to the SE, where the good intrabasement conductor seems to terminate. This interpretation would move the Rába Line from 62 to 70 km on regional line C. Balla et al. (1990) identified the Rába Line as a "dislocation zone" on the MK-1 seismic line. This boundary, however, does not seem visible on the seismic data (Plate 3). Balla et al. (1990) further interpreted the Rába Line as a major subvertical to vertical fault crossing the whole crust offsetting even the Moho (Fig. 4.18) based on a rather ambiguous gravity model calculation. Note also on Fig. 4.18 that the Mihályi high and its surroundings are erroneously classified as Penninic units ignoring the detailed petrological studies and age dates of Balázs (1971), Árkai et al. (1987) and Árkai and Balogh (1989).

Because of the considerable confusion surrounding the Rába Line outlined above, this fault will be redefined in Chapter 5 (cf. Tari and Horváth, 1992).

Strike-slip faults in the Transdanubian Central Range have been known since the beginning of the century (e.g. Lóczy, 1917). Telegdi-Roth (1935) recognized one of the most spectacular wrench faults running between Bakonybél and Várpalota, which is now called the Telegdi-Roth Line (Fig. 4.19). Systematic investigation of strike-slip features in the Bakony Mts. was carried out by Mészáros (1982, 1983, 1985, 1986) which he mapped on a 1:10,000 scale. According to Mészáros (1983), two major phases of wrench faulting can be distinguished, a Cretaceous (at the end of the Neocomian) phase and a Middle Miocene (Badenian and Sarmatian). All of the Miocene faults are right-lateral strike-slip faults with offsets up to 5 km. The Cretaceous set of wrench faults is less defined and is characterized by both dextral and sinistral faults. Note on Fig. 4.19 that Mészáros (1983) shows a major N-S trending dextral fault of Cretaceous age between Noszlop and Szög to explain the offset of the first-order syncline of the Bakony Mts. (cf. Fig. 4.13).
Fig. 4.18. Crustal structure along the MK-1 reflection seismic line (Balla et al., 1990). For location see Fig. 4.3.
The 4.7 km right-lateral offset on the Telegdi-Roth Line (Fig. 4.19) was determined by displaced Mesozoic terrains by Mészáros (1983). Originally, Telegdi-Roth (1935) thought that this strike-slip fault was essentially of Cretaceous age; however, allowing for displaced Eocene rocks, he supposed its reactivation in post-Eocene times. Kókay (1976) regarded all movements along this fault as Miocene in age. Tari (1991) proposed the reactivation of a WNW-trending Cretaceous fault set during the Middle Miocene based on a block rotation model. Tari (1991) further suggested that regardless of the age of these faults they all may be detached at mid-crustal depths (Fig. 4.20). This detachment surface probably emanated from the basal decollement of an Eoalpine nappe system.

The Telegdi-Roth Line and other NW-trending strike-slip faults shown in Fig. 4.19 can be indeed followed into the subsurface of the Danube Basin and their character and age are documented in Chapter 5.

In the northeastern part of the Transdanubian Central Range, other major E-trending strike-slip faults were described recently by Balla and Dudko (1989). These faults shown in Fig. 4.13 display both sinistral and dextral offsets. The Vértessomló-Nagykovácsi fault, for example, has an apparent sinistral strike-slip along its western part, while it shows 14 km of dextral shear on its eastern end. The dextral strike-slip seems to be younger than the sinistral according to Dudko (1992).

4.10.2 LATE MIOCENE-RECENT POST-RIFT TECTONICS

In the major Neogene basins of the area, post-rift thermal subsidence started at the Badenian/Sarmatian boundary. Very thick (up to 5 km), structurally undisturbed successions were deposited until the Quaternary. During the Quaternary, however, uplift of previously subsiding areas became common, heralding the inversion of the Pannonian Basin.

The thickest post-rift sedimentary fill can be found in the center of the Danube Basin
Fig. 4.20. Block diagram illustrating the proposed allochthonous structure of the Bakony Mts. (Tari, 1991). It was suggested that the base of the Eoalpine nappe system (heavy line) acted as the required basal detachment surface for the rotation of upper-crustal rigid blocks. Since its depth (10 km on average) is close to the inferred brittle-ductile transition during the Middle Miocene, the regional left-lateral shear might have been accommodated by creep along this surface (shaded) in the lower crust.
near the Hungarian/Slovakian border (cf. regional line F in Chapter 3). The Bösárcány-1 well (Fig. 4.3) penetrated the thickest known Sarmatian strata (630 m) of the whole Pannonian Basin, even though the well was drilled on the flank of the basin (Körössy, 1985).

Interestingly enough, the very center of the basin (Győr subbasin) still subsides synchronously with the uplift of the basin flanks (e.g. Rónai, 1986). Aside from regional line F, it is best illustrated by the present-day elevations of the Pleistocene river terraces of the Danube (Fig. 4.21). These terraces are correlated with the major interglacial periods of the last ice age which gives a good age-control (Pécsi, 1959; Funk, 1966). The vertically exaggerated section following the trace of the Danube (Fig. 4.21) clearly shows the ongoing uplift of the Transdanubian Central Range coeval with the subsidence of the Little Hungarian Plain (Danube Basin) and the Great Hungarian Plain. A possible explanation of this phenomenon is offered in Chapter 5.

4.10.3 MAGMATISM

Neogene magmatism is subdivided into two distinct periods in the NW Pannonian Basin. The Middle Miocene intermediate volcanism shows well-defined eruptive centers beneath the Styrian Basin (e.g. Gleichenberg) and the Danube Basin (e.g. Pásztori, Fig. 4.3).

Pliocene to Pleistocene mafic volcanism, however, formed numerous but small eruptive centers scattered evenly in the NW Pannonian Basin (Fig. 4.1). In addition to the basalt lavas, much tuff material can be found.

Körössy (1985) described the volcano of Pásztori (Fig. 4.3) encountered in several boreholes. The wells reached the top of this stratovolcano beneath the thick post-rift strata and found trachytic lavas and tuffs of Middle Miocene (Badenian) age. The volcanic activity continued until the lowermost Pannonian (see Fig. 4.11, Jámbor, 1989).
Fig. 4.21. Pleistocene uplift and subsidence of Danube river terraces, simplified after Pécsi (1959). For location see Fig. 4.3.
Some sixty-five occurrences of basalt lavas and tuffs are known in the Hungarian part of the NW Pannonian Basin (Jámbor et al., 1981). The major volcanic centers are shown in Fig. 4.1. Five periods of volcanic activity are distinguished based on stratigraphic and K/Ar age determinations ranging from the uppermost Late Miocene to the Early Pleistocene. The basalts have a strong alkaline character (Szabó et al., 1992). Frequent lherzolite inclusions indicate an upper mantle source of the magmatism (e.g. Embey-Isztin et al., 1990). The role of these "subsequent" basalt lavas and their relation to the Miocene extensional evolution of the Danube Basin are not understood.
CHAPTER 5
ALPINE TECTONICS OF THE NW PANNONIAN BASIN

In this chapter many of the problems outlined in Chapter 4 are addressed using primarily subsurface information (reflection seismic and well data). While regional sections A-F in Chapter 3 were intended to give only a general perspective of geology of NW Pannonian Basin, here more specific details and seismic illustrations will be given.

Fig. 5.1a shows the geology of the junction area of the Alps-Carpathians-Pannonian Basin. This region was subdivided into six areas (Fig. 5.1b) and thus the descriptions of the individual Alpine stages are organized in a spatial and temporal framework (Fig. 5.2). In contrast to the previous chapters, the discussion of the structural interpretation begins with the present-day situation and goes backwards in time. This description, however, ends close to the Cretaceous/Jurassic boundary with the Eoalpine stage (Fig. 5.2), since the presently available seismic and well data do not yield any direct information on the previous pre-Alpine and early Alpine evolution of the area.

The area of NW Hungary (Fig. 5.3) covered by seismic reflection data is subdivided into twelve subareas. The description of individual seismic lines is organized in relation to these subareas. To emphasize the regional implications of this chapter, a regional cross-section was constructed based on depth-converted reflection seismic lines.

5.1 ORGANIZATION AND PRESENTATION OF SEISMIC DOCUMENTATION

Less than one-fifth of all available seismic data is here shown on panels (Panels 4-15). Moreover, since these sections involve different vintages of very different quality and reproduction, half of them are shown only as line drawing interpretations. All the sections are migrated and are 1:1 for a velocity of 5,000 m/s (16,400 ft/s). Datum is 100 m (328 ft) above sea level.
Fig. 5.1a. Major tectonic units of the Alps-Carpathians-Pannonian Basin junction area.
Fig. 5.1b. Index map of the Alps-Carpathians-Pannonian Basin junction area.
<table>
<thead>
<tr>
<th>TIME IN MA</th>
<th>AREAS</th>
<th>V</th>
<th>S</th>
<th>D</th>
<th>Z</th>
<th>B</th>
<th>G</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Vienna Basin and Northern Calcareous Alps</td>
<td>Styrian Basin and South Burgenland basement high</td>
<td>Danube Basin (Hungary/Slovakia)</td>
<td>Zala Basin and Balaton zone</td>
<td>Bakony, Vértes and Velence Mts., Balaton Highland</td>
<td>Gerecse and Buda Mts.</td>
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<td>-10</td>
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<td>post-riptide subsidence</td>
<td>post-riptide subsidence</td>
<td>inversion</td>
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<td>inversion</td>
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<td>wide-riptide extension</td>
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<td>core complex ext.</td>
<td>wide-riptide extension</td>
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<td>escape tectonics</td>
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<td>uplift and erosion</td>
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<td>Maastrichtian</td>
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<td>extensional (?) basin</td>
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<td>overthrusting</td>
<td>forebulge erosion</td>
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<td>forebulge erosion</td>
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<tr>
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<td>no sedimentary record overthrusting</td>
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</tr>
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<tr>
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<td>Aptian</td>
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<td>passive margin</td>
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<td>passive margin</td>
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<tr>
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<td>Hauterivian</td>
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<td>passive margin</td>
<td>passive margin</td>
<td>passive margin</td>
<td>passive margin</td>
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Fig. 5.2. Summary of Alpine tectonics (Cretaceous to Recent) in the Alps-Carpathians-Pannonian Basin junction area.
Fig. 5.3. Seismic and well data available for this study in the NW Pannonian Basin.
The seismic and well data cover the numerous subbasins and basement highs of the Hungarian part of the Danube and Zala Basins (Fig. 5.3). The names of seismic sections and wells are shown in Fig. 5.4a,c (see also Map 1) and listed in Appendix D.

5.1.1 DEFINITION OF SUBAREAS IN NW HUNGARY

The systematic description of reflection seismic profiles follows the subdivision shown on Fig. 5.4a. Single seismic lines are identified by the letter of the subarea and a number. Other seismic illustrations described in Chapter 3 are also shown on Fig. 5.4a.

Subarea C (Csapod) is the northwestern part of the Hungarian Danube Basin, including the Nagycenk and Csapod Basins, the Pinnye high.

Subarea M (Mihályi) is the Mihályi basement high in the central part of the Hungarian Danube Basin, between the Csapod and Kenyeri Basins.

Subarea P (Pásztori) is the Pásztori paleovolcano in the central part of the Hungarian Danube Basin, between the Mihályi high and the Győr and Kenyeri Basins.

Subarea N (Nagyigmánd) is the Pér-Nagyigmánd area in the northeastern part of the Hungarian Danube Basin, between the Győr Basin and the Transdanubian Central Range.

Subarea G (Györszemere) area of the northeastern part of the Hungarian Danube Basin, located between the Pásztori paleovolcano and the Transdanubian Central Range.

Subarea V (Vaszar) is the Vaszar-Takácsi basement promontory of the Transdanubian Central Range in the area in the eastern part of the Hungarian Danube Basin to the SE of the Pásztori paleovolcano.

Subarea D (Dabrony) is the Dabrony-Kenyeri Basin in the eastern part of the Hungarian Danube Basin, between the Mihályi high and the Transdanubian Central Range.

Subarea K (Káld) is the Káld-Ukk basement promontory of the Transdanubian Central Range in the southeastern part of the Hungarian Danube Basin, to the SW of the Dabrony Basin.
Fig. 5.4a. Index map of subareas and seismic data utilized in this thesis. The location of this map is shown in Fig. 5.3
Fig. 5.4b. Index map of seismic illustrations in this study. The location of this map is shown in Fig. 5.3.
Fig. 5.4c. Index map of wells utilized in this study. The location of this map is shown in Fig. 5.3. In the southern part (Zala Basin) there are many more wells than shown.
Fig. 5.4d. Tectonic map of NW Hungary based on reflection seismic data interpretation (this work) and well control. Compare with published maps of others, see e.g. Fig. 4.3.
Subarea I (Ikervár) is the Ikervár-Nemeskölta-Egyházasrádóc basement high and the Hungarian part of the South-Burgenland high in the southwestern part of the Hungarian Danube Basin, to the SW of the Mihályi high.

Subarea A (Andráshida) is the Andráshida-Nádasd-Geresd-Nagytilaj basement promontory of the Transdanubian Central Range between the southern part of the Hungarian Danube Basin and the northern part of the Zala Basin.

Subarea S (Szentgotthárd) is the Szentgotthárd basement high as the Hungarian part of the South-Burgenland high and the Óriszentpéter Basin of the northwestern part of the Zala Basin.

Subarea Z (Zalatárnok) is the Nagylengyel basement promontory of the Transdanubian Central Range and the Zalatárnok-Bak trough in the central part of the Zala Basin.

5.1.2 STRATIGRAPHY IN TERMS OF SEISMIC CHARACTERISTICS

Fig. 5.5 is a simplified summary of the stratigraphy of the Hungarian part of the NW Pannonian Basin, based on the detailed lithostratigraphy described in Chapter 4. While the thickness data are well known for the upper 10 km in this composite section, the thickness relations are poorly constrained for the Paleozoic of the Austroalpine units and for the Mesozoic of the Penninic unit.

Interval velocities of the major units were compiled based on well-shots and reported interval velocities in several seismic surveys. Fig. 5.5 also shows the interval velocity which was adopted for the depth-conversion of selected seismic sections.

The seismic horizons were picked as in the case of the regional sections of Chapter 3 (see Fig. 3.3) with some important additions (Fig. 5.6). These latter include an intra-syn-rift unconformity, an intra-Sernenian local unconformity and an unconformity between the Middle and Lower Cretaceous. Note that the Hercynian unconformity at about 10 km
<table>
<thead>
<tr>
<th>STAGES</th>
<th>TECTONIC</th>
<th>SIMPLIFIED LITHOLOGY</th>
<th>AVERAGE THICKNESS (M) (RANGE)</th>
<th>INTERVAL VELOCITIES (M/S) (RANGE)</th>
<th>ADOPTED FOR DEPTH CONVERSION</th>
<th>SEISMIC HORIZONS</th>
</tr>
</thead>
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<td>-1</td>
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<td>NEOGALPINE</td>
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<td>2585 (1920-3250)</td>
<td>2500</td>
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</tr>
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<td></td>
<td></td>
<td>500 (150-1000)</td>
<td>3185 (2450-3920)</td>
<td>3100</td>
<td></td>
</tr>
<tr>
<td>-3</td>
<td></td>
<td>MIDDLE ALPINE</td>
<td>600 (170-1100)</td>
<td>3450 (2900-4500)</td>
<td>3500</td>
<td></td>
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<td>-4</td>
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<td></td>
<td></td>
<td>800 (480-1200)</td>
<td>4400 (3750-5050)</td>
<td>4400</td>
<td></td>
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<td>500 (320-850)</td>
<td></td>
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<td>-7</td>
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<td>500 (480-950)</td>
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<td>300 (180-450)</td>
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<td>-9</td>
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<td></td>
<td>4500 (3050-6500)</td>
<td>5850 (4500-6550)</td>
<td>6000</td>
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<td>-11</td>
<td></td>
<td></td>
<td>2000 (2000-4000?)</td>
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<td>-12</td>
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<td></td>
<td>1000 (500-4000?)</td>
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<tr>
<td>-13</td>
<td></td>
<td></td>
<td>1500 (1000-4000?)</td>
<td>6050 (4700-6800)</td>
<td>6100</td>
<td></td>
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</tbody>
</table>

Fig. 5.5. Lithology and seismic characteristics in the NW Pannonian Basin.
Fig. 5.6. Legend of seismic sections (Panels 4-15) in the northwestern part (upper) and southeastern part (lower) of the NW Pannonian Basin, Hungary. Dotted pattern overlying the normal fault on the upper legend refers to clastic fans. Note that the Early Alpine unconformity is near the top of Neocomian.
(Fig. 5.5) is not correlated on the reflection seismic sections since it does not have a
distinct seismic signature (!). However, for the very first time two regional Eoalpine
tectonic contacts were correlated based on seismic expression: the top of the Lower and
Middle Austroalpine units. The top of the Penninic unit has a very marked seismic
expression and it was correlated across the western part of the basin.

In order to present a legible display of the legends I use different versions for the
northwestern and southeastern subareas of the study area (Fig. 5.6).

5.2 INTERPRETATION OF REFLECTION SEISMIC LINES

The seismic illustrations are shown in panels (Panels 4-15). On any one panel three
seismic sections are shown, commonly two dip-sections and a strike section connecting
them. In the following I briefly sum up the most important elements of the interpreted
seismic sections. A new approach to the seismic interpretation of intra-basement structures
is introduced in subchapter 5.2.5. The observations of all the following subchapters below
on the structure and stratigraphy of each subarea are summarized in chronological order in
subchapter 5.4 (see also Fig. 5.2).

5.2.1 AREA C (CSAPOD)

The structure of Area C in the northwestern part of the basin (Fig. 5.4) is
documented by four panels (Panels 4-7). In this area the Neogene basin fill displays a
general monoclinal dip to the E. While the post-rift Pannonian succession covers all the
pre-Tertiary basement structures, the syn-rift Middle Miocene (Karpatian-Badenian) can
be found only in local subbasins, between basement highs. The two prominent basement
highs (Bük-Pinnye and Mihályi-Mosonszentjános) of area C strike to the NE-N and
delimit the Csapod subbasin in between (Fig. 5.3).

Starting from the SW, consecutive dip-oriented (i.e. NW-SE) seismic sections reveal
the gradual deepening and widening of the Csapod subbasin due to increasing normal offset along a major detachment fault on its northwestern flank (see sections C1, C2, C3, C4, C5, C6, C7 and C10). This is clearly a low-angle fault as it flattens at depth, therefore it is regarded as a listric normal fault sensu Bally et al. (1981). The fault corresponds to the Alpokalja Line of Fülöp (1989, 192 p.) although he also called it the Répce Line after the nearby Répce River (Fülöp, 1990; 78 p.). The seismic shows that this fault is not a strike-slip fault (cf. criteria given by Harding, 1985; 1990) as considered by many (see Fig. 4.12). In the following I will refer to this low-angle normal fault as the Répce fault (Fig. 5.4d). Note that in this work I use the terms low-angle and detachment fault interchangeably.

The fault plane itself can be traced between the terminations of more or less coherent SE-dipping basement reflectors of the Pinnye high and the overlying chaotic seismic facies that may correspond to coarse-grained clastics. This later facies represents alluvial talus sediments deposited synchronously with the initial activity of the fault. On section C6 this facies was penetrated by the Csapod-1 well which indeed drilled about 500 m thick Karpatian succession of conglomerates and breccias (Kőrössy, 1987).

The Badenian syn-rift fill of the Csapod subbasin is just slightly asymmetric and displays little or negligible growth. This indicates that much of the normal faulting occurred right at the beginning of rifting, i.e. during the Karpatian. Strikingly similar seismic examples of analogue basins were published from the Basin and Range province by Effimov and Pinezich (1981) and from the Newark Basin by Costain and Coruh (1989).

Interestingly enough, coherent basement reflector packages below the Mihályi high display a roll-over anticline (see sections C3, C4, C5, C6 and C7) apparently associated with the large normal offset on the Répce detachment fault. Note that in some cases (sections C3, C4, C5, C6, C7 and C10) the normal offset can be estimated by restoring displaced prominent basement reflectors in the Bük-Pinnye and Mihályi-Mosonszentjános
highs. Such a reconstruction gives normal offsets (heave) of about 4-6 km for the horizontal component, with relatively small error bars of about 0.5 km. The heaves show an overall increase to the NE along the Répce fault, with a local maximum occurring at sections C5 and C6 (see below).

An intrusive body was interpreted on sections C1 and C17. A corresponding magnetic anomaly has been known for a long time and Posgay (1966) attributed it to a Paleozoic intrusive body at 4 km depth. The nearby Bük wells did not reach this body but the seismic evidence shows that this intrusion is definitely post-Lower Cretaceous, since it had a profound effect on Penninic reflector packages. Furthermore, this intrusion seems to be post-detachment faulting (i.e. post-Karpatian), since it bends the tectonic top of the Penninic unit as well. If this is really the case then the intrusion represents part of the widespread trachytic volcanism of the Pásztori volcano (see below).

Another fine example of magmatic activity can be seen on section C5, where on the southeastern flank of the Mihályi high two small mushroom-shaped bodies are shown to intrude the Pannonian succession. These are apparently basalt laccoliths related to the Pliocene Tapolca Basalt (Fig. 4.11).

Note that within the basement the Répce detachment fault shows up with prominent fault-plane reflectors (e.g. sections C6, C7 and C10). Comparable fault plane reflectors of similar detachment fault planes were reported from Utah (e.g. Allmendinger et al., 1983 and von Tisch et al., 1985) and Arizona (e.g. Frost and Okaya, 1986).

Since many intra-basement reflecting horizons could be correlated with considerable confidence in area C, I have attempted to map certain basement units based on their seismic character. In a trial-and-error procedure I came up with an interpretation which fits the available well control in the area with the following assumptions:

a) The very low-grade to low-grade metamorphics drilled in numerous wells on the Bük and Mihályi highs (Fig. 5.4c) are part of the Upper Austroalpine nappe system. This
interpretation is supported by the studies of Árkai et al. (1987) and Árkai and Balogh (1989). However, it contrasts with the subdivision of Fülop (1989, 1990), who separated these successions, and with the idea of Balla et al. (1990), who regarded these units as Penninic (see Fig. 4.18).

b) The Middle Austroalpine is present in the basement of area C and the metamorphics (paragneiss and micaschist) encountered in the Pi-1,-2; Mos-1,-2 and M-4 wells (Fig. 5.4c) may be part of this Austroalpine unit. This is indirectly supported by Deák (1981; 34. p), who reported clasts of very high-grade metamorphics from the Csapod well (section C6), that appear to be correlatable with the Middle Austroalpine Sieggraben series (for details see Appendix A). This assumption contradicts the traditional view (e.g. Fülöp, 1990) which regards all of the medium-grade metamorphics encountered in the above wells as Lower Austroalpine correlatable with the Grobgneiss and Wechsel series of the Sopron area (Fig. 4.4).

c) The Lower Austroalpine probably subcrops in the pre-Tertiary basement only in the northernmost part of NW Hungary, in the Msz-1,-2 and Raj-1 wells (Fig. 5.4c). This is supported by the presence of a Permian(?)-Triassic red sandstone succession encountered in the Msz(Mosonszolnok)-1 well which can be correlated with the Lower Austroalpine Semmering series (Körössy, 1987).

In addition to these assumptions a very important geometric constraint is provided by the Penninic succession which has a very distinct, highly reflective seismic expression and a well-defined top (see e.g. Fig. 3.7a).

Since the Upper Austroalpine unit is lithologically markedly different from the Middle Austroalpine (very low-grade to low-grade versus medium-grade metamorphics) their contact is interpreted to correspond to a pronounced change in reflectivity (see e.g. section C2). The Upper Austroalpine unit is characterized by short, but strong reflectors in contrast to the underlying Middle Austroalpine which has a mostly transparent character.
In the southwestern part of area C the Upper Austroalpine can be found right on top of the Penninic with a fault contact (sections C1, C2 and C17). This relationship was indeed observed along strike in the Eisenberg Mountains (Fig. 5.2) where the Upper Austroalpine Hannersdorf series have a poorly understood tectonic contact with the Penninic succession (e.g. Pahr, 1980; Schmidt et al., 1984; Tollmann, 1989).

On sections C1, C2 and C3 another detachment fault can be interpreted to the NW of the Bükk high forming a smaller syn-rift graben. I refer to this detachment fault as the *Ikva fault* after the nearby river and to the associated basin as the Zsfrá subbasin (Fig. 5.4d). The Ikva fault is detached on top of the Penninic in the above mentioned sections. Farther to the NE the fault shows gradually decreasing normal offset and flattens out close to base of the inferred Middle Austroalpine unit (sections C5 and C6).

While in the southwestern part of area C (sections C1 and C2) the Répce fault seems to flatten close to or into the Upper/Middle Austroalpine boundary (see strike section M18), it apparently ramps down to deeper stratigraphic levels along strike, i.e. to the NE (sections C3, C4 and C5).

The Middle/Lower Austroalpine units cannot be distinguished based on seismic facies since their lithologies are quite similar (i.e. dominantly medium-grade metamorphics). A somewhat arbitrary boundary is placed at a strong reflection doublet which shows up on some of the sections (see section C17) and which may be associated with fault plane reflectors. Thus the Middle Austroalpine unit interpreted in this manner is about 900-1000 ms thick, which means about 3 km thickness assuming 6000 m/s interval velocity. Alternatively the strong reflectors may originate from a thin (200-300 m) layer of the Permomesozoic cover sequence embedded within the Lower Austroalpine unit.

As it is shown on strike section C20 the Répce fault plane displays a synform with a local low close to section C5. Since the dip section C5 reveals the deepest part of the Csapod basin as due to a local maximum in the normal offset of the Répce fault, the small
flat within the Lower Austroalpine is perhaps associated with a very "efficient" detachment level. Note that the pronounced synclinal feature is even more remarkably displayed on section M18. Considering the dip lines (C4, C5 and C6) the Répce fault has a "spoon" shape in the basement with maximum displacement along the long axis of the spoon. Interestingly, the Répce fault plane climbs up in terms of physical depth farther to the NE, but it ramps down stratigraphically into the Lower Austroalpine (sections C20 and M18).

5.2.2 AREA M (MIHÁLYI)

The structure of Area M in the northwestern part of the basin (Fig. 5.4) is documented by one panel (Panel 8). In this area the prominent basement high (Mihályi-Mosonszentjános) strikes to the NNE and it is flanked by the Csapod subbasin to the NW and the Kenyeri and Győr subbasins to the SE (Fig. 5.3).

The Mihályi structure becomes a pronounced basement high to the NE as the onlapping syn-rift and post-rift reflectors indicate (sections M1 and M3). It is a fault bounded high and the normal offset on the boundary fault at the southeastern flank gradually increases to the NE. This fault is identified as the enigmatic Rába Line in the sense of Horváth et al. (1987) and Rumpler and Horváth (1988), and in the following I refer to this low-angle normal fault (cf. Figs. 3.12 and 4.17) as the Rába fault (Fig. 5.4d). This fault, similarly to the Répce fault, is not a strike-slip fault (cf. criteria given by Harding, 1985; 1990).

The Rába fault can be traced to the SW to the northwestern flank of the Ikervár high where it has zero normal offset and it coincides with a SE-dipping basement structure that appears to be a pre-existing thrust plane (Fig. 5.4d, cf. sections M1 and I16). The dip of the Rába fault in the Mihályi area is generally steeper (40-50°) than the Répce fault and cuts the Eoalpine overthrust contacts. The fault plane, however, shows a progressively more low-angle character with increasing normal offset to the NE.
Like the Répce fault, seismic facies suggests the presence of thick alluvial fans above the fault plane shed from the adjacent high. Middle Miocene coarse conglomerates were indeed found in well M-28 (see section M3; Körössy, 1987). Beneath the Kenyeri subbasin a smaller normal fault block can be interpreted on several sections, indicating the along-strike branching of the Rába fault (e.g. section M3).

As the Rába fault becomes more low-angle to the NE it is harder to pick the updip termination of the fault plane on the basement high. Another problem relates to the fact that the prolonged syn-rift and partly post-rift subaerial exposure of the Mihályi basement rounded off the contour of the high, and it certainly removed a portion of the updip end of the fault. The downdip geometry of the Rába fault is harder to assess when compared with the Répce fault, because the available industry lines do not record the critical depth of about 5-6 s TWT time (~15-20 km), where the flattening of this fault could be expected (cf. Plate 3).

The internal geometry of the Mihályi high reveals a roll-over anticline due to the movement on the Répce fault (section M3). Since the Rába fault offsets this anticline, the latter fault seems to postdate the Répce detachment fault.

5.2.3 AREA P (PÁSZTORI)

According to the three wells drilled in the vicinity of Pásztori and the large magnetic anomaly in the same region, a large stratovolcano separates the Kenyeri and Győr subbasins to the SE of the Mihályi high (Figs. 5.3 and 5.4d). The stratovolcano produced a >1727 m thick volcano-sedimentary succession penetrated by the Pá-1 well (Körössy, 1987). Thick claymarl intercalations occur within the volcanic sequence and suggest submarine volcanism. The trachytic-andesitic volcanism might have begun as early as the Karpatian but definitely continued until the Sarmatian and possibly into the Early Pannonian (see Fig. 4.11).
By Sarmatian the central part of the Pásztori volcano was above sea-level as the onlapping reflectors show (see section P26 in Panel 9). Volcanism during this time must have been analogous to the present-day Stromboli of the Lipari Islands.

The volcanics have a distinct seismic reflection pattern: wavy, short and strong reflections (section P26). This facies shows a gradual transition towards the SE into a dominantly sedimentary syn-rift succession (sections G2, G3). The top syn-rift unconformity can be correlated in area P with only little confidence.

The top of the pre-Tertiary basement, however, can sometimes be picked even below the Pásztori volcano as a package of strong reflectors (section P26). The maximum thickness of the volcanic succession is about 1500 ms equivalent to about 2300 m (at velocity 3000 m/s). Adequate mapping of intra-basement structures is difficult if not impossible, although the Eoalpine nappe units are probably present (see coherent intra-basement reflectors in section P26). Körössy (1987) tentatively placed his Rába "Line" to the NW of the Pásztori volcano. I have identified the Rába fault as a poorly defined low-angle normal fault on section G2, beneath the Bösárkány(Bő)-1 well.

5.2.4 AREA N (NAGYIGMÁND)

Subarea N is covered by a number of seismic reflection sections (cf. Fig. 5.3 and 5.4) whose quality unfortunately is relatively poor since these data were gathered during the seventies. Therefore no seismic illustration is shown from this subarea.

In this area the Neogene basin fill displays a general monoclinal dip to the NW above the Mesozoic basement (see regional section F in Panel 1 and also a crustal section shown in Plate 6). Interestingly enough, in spite of the northwesterly dip within the post-rift Pannonian succession, a southeastward prograding pattern can be observed which suggests a very recent basinward tilt of the basin margin. This feature is very characteristic for much of the southeastern flank of the Hungarian Danube Basin, as will be discussed in
subchapter 5. 4.1.1 (see also subareas G, D, I and A below).

5.2.5 AREA G (GYÖRSZEMERE)

Aspects of this subarea were previously mentioned in the description of the Pásztori volcano. Two dip profiles (sections G2 and G3 in Panel 9) were reproduced to show the structural characteristics in this part of the Danube Basin.

Again the whole post-rift sequence with its internal southeasterly progradation displays a pronounced tilt to the NW. Note that even the youngest Pannonian reflectors show a near-surface truncation, suggesting the very young differential uplift of the Bakony Mts. to the SE. Unfortunately these youngest Pannonian reflectors cannot be followed updip beyond a certain depth because the uppermost 300 ms of the seismic data was muted - a common practice in the Hungarian oil industry.

Within the post-rift succession, flower structures can be seen. They show up also on some of the other lines trending to the NE. Within the syn-rift sequence which displays a clear growth towards the basin, a number of internal unconformities can be interpreted. These unconformities were studied in detail by Mattick et al. (1989). In this study I have differentiated only one unconformity within the syn-rift succession (sections G2 and G3). This boundary tops a peculiar unit which thickens towards the SE. According to Mattick et al. (1989) this unit is made up of the Middle Miocene (Karpatian?) continental red beds and conglomerates. This is the basis for their classification as part of the syn-rift unit. Alternatively, these continental clastics could also represent the Upper Oligocene to Lower Miocene Csatka Gravel. If so, the unit belongs to the uppermost sequence of the Paleogene Basin.

In any case the Tertiary sequence was deposited on top of a pre-Senonian basement that is a very well-defined surface in almost all of the sections, as it is characterized by pronounced relief differences (50-200 ms TWT time). The consistently SE-dipping intra-
basement reflectors display a close relationship to the paleomorphology of the basement (e.g. sections G2 and G3).

At this point it is important to note that even though the presence of intra-basement reflectors is undeniable in the study area as compared to other regions of the Pannonian Basin (see Chapter 3), mapping them is still difficult. This is because the strong basement reflectors interpreted as overthrust planes (see Fig. 3.14) do not show up on every consecutive seismic section. Thus, the mapping of Eoalpine thrusts based on these intra-basement reflectors had very limited success (cf. Mattick et al., 1989), and my attempts using conventional mapping techniques have also failed.

An important observation, however, offers a new approach to interpretation. I found what appear to be systematic cuestas (or hogbacks) below the well-mappable top of pre-Senonian surface. The relief of these cuestas is in the order of 50-200 ms TWT time, or approximately 100-400 m. The cuestas are very well traceable on consecutive dip sections.

I suggest that the morphology of the pre-Senonian basement is the result of strong differential erosion that took place before the transgression of the Senonian Sea. Karst-formation and bauxite deposition indeed occurred between the Cenomanian and the Santonian (see Chapter 4). In the Mesozoic column of the Bakony Mts. (Fig. 5.7), the most resistant formation seems to be the Hauptdolomit. The overlying Kössen beds and the underlying VeszpréM Marl are expected to be recessive because of their lithology. Thus Fig. 5.7 illustrates the potential resistance to long-term (~5 Ma in this case) subaerial exposure.

This interpretation can be readily verified since the erosional highs due to the resistant beds (either of Hauptdolomite, Dachstein Limestone or Megyehegy Dolomite) were drilled along strike (e.g. Gysz-3 and Tét-2 in this area). Because I found a very good match between the interpreted lithology and well data, I used this procedure to map the top of the pre-Senonian basement in areas that lack well control (see Fig. 5.4d). Certain
Fig. 5.7. Summary diagram of erosional resistance of the Triassic in the Bakony Mts. Potential décollement levels are also shown, see discussion in subchapter 5.4.3.2.
Paleozoic units in the deeper part of the Upper Austroalpine (see below the case of the Ikervár high in area I) can also be mapped using the same technique.

As an example, the intra-basement structure of section D2 may be interpreted as follows. The Gysz-3 well penetrated Middle Triassic Megyehegy Dolomite and immediately below Paleozoic schists. Teleki et al. (1989) interpreted this contact as a thrust, although the stratigraphic omission would rather indicate normal faulting. Local omission of tectonostratigraphy, however, does occur along the well-known Litér thrust in the Balaton Highland and therefore the thrust interpretation is accepted. The prominent cuesta 2 km updip from the Gysz-3 well consists of Hauptdolomite, assuming a 30° SE dip and the average thickness of the Triassic succession (Fig. 5.5). Beyond that a major thrust plane follows which puts Paleozoic on top of Upper Triassic. This is confirmed by the nearby Tét-2 well which could be projected into section D2 along strike from a distance of 1.5 km, 5.5 km updip from the Tét-1 well (Fig. 5.4c). The other cuesta 2 km downdip of the Gysz-3 well consists of Paleozoic sandstoneschists and greenschists since this very pronounced paleotopographic feature can be followed along strike to the SW for about 55 km(!) to the Ikervár high where a number of wells drilled into it (Fig. 5.4d).

Note that the magnetotelluric Rába Line of many workers (e.g. Balla et al., 1990) is associated with the thrust fault subcropping on the pre-Tertiary surface between Gysz-3 and Tét-1 wells on section G2. The Paleozoic of this thrust sheet is at least 2-3 km thick and therefore must involve the deeper stratigraphic levels of the very-low grade Paleozoic strata. Since Ádám et al. (1990) recently identified the good conductivity zone in outcrops as being Paleozoic graphitic shales of the Gailtal area in the Eastern Alps (Fig. 2.8), they proposed the same origin for the widespread Transdanubian anomaly as well. The age of the graphitic shales in the Gailtal area is Ordovician-Silurian (Schönlau, 1979), and I infer the presence of the same stratigraphic level in the subsurface of the Bakony Mts. I further speculate that this graphitic lithology not known from outcrops in Hungary also provided
the actual décollement surface for this particular thrust sheet.

Some 7 km downdip from the Gysz-3 well on section G2 another cuesta may correspond to another thrust plane indirectly supported by the minor extensional reactivation of the same surface on section G3. This thrust plane is correlated with the incipient Rába fault in section M1 (Fig. 5.4d) and separates very low-grade to low-grade Paleozoic units of the Upper Austroalpine.

Farther downdip, towards the center of the basin the seismic data deteriorate beneath the thick syn-rift volcanosedimentary blanket, preventing correlation of topography on top of the pre-Tertiary basement.

Finally, it is important to emphasize that only the map view of the pre-Tertiary basement can be determined using the above described correlation technique. The geometry of thrust sheets deep within the basement besides the overall SE dip (20-40°, or 30-50° restoring the Neogene differential subsidence) cannot be constrained with confidence because of the deterioration of data with depth (see e.g. sections G2 and G3) and becomes a matter of conceptual interpretation (see regional structure section below).

5.2.6 AREA V (VASZAR)

Similarly to area N the seismic reflection data in this area are also represented mostly by sections acquired during the seventies. Therefore I did not attempt to illustrate these lines on a panel.

One published seismic section, however, is considered fairly typical for the area, as shown on Fig. 3.14. The intra-basement thrust plane interpreted originally by Horváth and Rumpler (1984) was confirmed, as it coincides with one of the regionally mapped thrust contacts of this study (Fig. 5.4d). This mapping, however, disproved the presence of Penninic metamorphites below the thrust plane as it was proposed by the above authors.

The most important structural feature is a steeply dipping thrust which obliquely
cross-cuts earlier thrust planes (Fig. 5.4d). Therefore this thrust apparently emanated from a deeper décollement level (see e.g. Bally et al., 1966) and it is responsible for the local basement high of Takácsi, where a number of wells (Takácsi and Vaszar wells, Fig. 5.4c) drilled into the very low-grade Paleozoic basement beneath the Neogene. The age of this thrust is certainly pre-Neogene and, since it seems to be responsible for the erosional truncation and termination of the Senonian succession of area D (see below), its age is bracketed within the Paleogene. This distinctive NNW-verging thrust plane shows up on a number of consecutive sections, but it cannot be traced to the WSW beyond the Telegdi-Roth Line (Fig. 5.4d).

5.2.7 AREA D (DABRONY)

This area is characterized by the local thickening of the Miocene syn-rift sequence and the underlying Senonian succession as well. The Dabrony subbasin is delimited by a number of NW-trending steep normal faults to the SW and by the Telegdi-Roth right-lateral strike-slip fault zone to the NE (Fig. 5.4d). To the NW the Dabrony subbasin merges into the Kenyeri subbasin. Seismic illustrations of this key area are shown in Panels 10 and 11.

The Neogene succession displays the same overall geometry as in the previously discussed N, G and V areas. The northwestern end of dip lines (sections D9, D12, D13 and D14) show the thick Miocene syn-rift fill of the Kenyeri subbasin controlled by the Rába fault. Another dip line, section D4 shown in Fig. 3.12, was briefly discussed earlier. The lower half of the syn-rift succession displays considerable growth (sections D9 and D12), but since no well was drilled in the center of the basin its exact age is unknown. An educated guess for this age is Karpatian, because the marked unconformity observed in the Kenyeri subbasin may be correlated updip into a similar intra-syn-rift unconformity in the shallower Dabrony subbasin. This latter unconformity shows up as a pronounced angular
discordance on the strike lines (sections D31 and D33). According to the well reports this un conformity separates Karpatian continental clastics from Badenian neritic clastics and carbonates (Körössy, 1987; Teleki et al., 1989). This particular age and the angular un conformity suggests that this stratigraphic boundary is correlatable with the well-known Styrian un conformity of the Styrian Basin (Fig. 5.1) in Austria (Stille, 1924; Friebe, 1991a,b; 1993). The correlation of this un conformity towards the basin center is problematic. The strike-lines (sections D31 and D33) show a fairly uniform thickness of the Karpatian sequence as opposed to the syn-extensional growth of the Badenian strata.

Multistage structural history can be outlined for the ~5 km wide zone of the Telegdi-Roth Line (Fig. 5.4d) as revealed by strike sections D31 and D33. This right-lateral fault zone was definitely active during the Middle Miocene and the transpressional inversion along previous normal faults is quite clear. Remarkably the 4.7 km of total dextral offset on the two branches of this wrench fault suggested by Mészáros (1983) based on map relations within the Bakony Mts (Fig. 4.19) is reconfirmed. The strain partitioning due to Cretaceous and Miocene strike-slip movements (cf. Telegdi-Roth, 1935; Kőkay, 1976, 1985; Mészáros, 1983; Tari; 1991) still cannot be differentiated. At any rate, the Telegdi-Roth fault zone cannot be traced across the Kenyeri subbasin and does not offset the crest of the Mihályi high (cf. Figs. 4.3b and 5.4d). This lends credibility to an earlier speculation of Tari (1991) that this and other similar strike-slip faults in the Bakony Mts. should be detached at depth on an Eoalpine thrust plane (see Fig. 4.20). The tectonic map in Fig. 5.4d shows a possible candidate for this detachment surface (see also area K below). The same explanation holds for the "still unexplained" termination of strike-slip faults at major thrust faults in the Balaton Highland (Mészáros, 1983; Fig. 4.19) which puzzled many (e.g. Dudko, 1991).

Similar detachment of simple Miocene normal faults on Eoalpine thrust planes at depth can also be observed, for example on strike section D31. Minor extensional
reactivation of Eoalpine thrust contacts can be seen on dip sections as well (sections D12 and D13).

The map view of Neogene structural elements based on seismic reflection data (Fig. 5.4d) excludes the interpretation of the Dabrony subbasin as a pull-apart basin (cf. Dudko, 1991a; Dudko et al., 1992). These authors proposed a Miocene pull-apart basin associated with a NE-trending left-lateral strike-slip fault, based solely on gravity anomalies and the stratigraphy of three wells in the area.

A striking difference in area D with the previously described areas is the presence of a thick (up to 1000 m) Senonian succession which was drilled in five wells in the Dabrony subbasin (Cell-Ny-1, Cell-1, Vi-1, Da-1 and Ncsd-1; Fig. 5.4c). The boundary between the Neogene and Senonian successions was correlated on the seismic based on a top reported from the Da-1 well (section D9). Note that this results in a top some 150 ms higher in the Vi-1 well (section D12) than the well documentation shows.

The Senonian sequence has a distinct seismic expression. The upper part, dominated by the Polány Marl, has a transparent character, whereas the strong reflectors immediately above the pre-Senonian basement correspond to a number of lithologies. Most importantly some reef-like build-ups can be interpreted on almost all the lines. These reefs(?) probably made up of the Ugod Limestone tend to be associated with the local topographic highs of the basement and can be correlated along strike.

Note that a locally developed intra-Senonian unconformity can be picked on a number of lines based on a prominent onlap surface (sections D9 and D13). This unconformity shows up in the upper, marly part of the Senonian sequence. Since it appears to be associated with a certain basement thrust unit, a minor compressional reactivation of that thrust during Senonian seems to be a plausible explanation for the appearance of this unconformity (see detailed discussion in 5.4.3.1).

Within the pre-Senonian basement itself a number of thrust sheets can be mapped.
Of particular importance is section D14, where even the cutoff geometry can be interpreted with reasonable confidence. The updip portion of section D14 deteriorates since it coincides with the Telegdi-Roth fault zone (Fig. 5.4d).

5.2.8 AREA K (KÁLD)

This area covers a topographic high of the pre-Tertiary basement with a relatively thin (500-1000 m) Neogene cover. From the moderate quality seismic data three illustrations are reproduced in Panel 12.

The southeastern part of the Káld basement high is covered dominantly by post-rift sediments (section K3), although the syn-rift succession is also present according to well control, but apparently in a very reduced thickness. The syn-rift strata fill mainly the topographic depressions of the basement (section K4).

The general SE dip of the intra-basement features is clear on all the lines. Three wells drilled on this basement high (Káld-1, Bor-1 and Mes-1; Fig. 5.4c) provided the constraints for the interpretation and correlation of thrust sheets. The Bor-1 and Mes-1 wells are particularly important, since Middle Triassic and Lower Triassic carbonates, respectively, were reported from them (Körössy, 1987).

On the northeastern end of strike section K11, the edge of the Dabrony subbasin can be seen (Fig. 5.4d). Close to the southwestern end of this section a steeply dipping, WNW-striking reverse fault was interpreted which could also be traced on parallel strike lines. This feature was already active before the Miocene and created a relief of about 300 ms (~400-500 m). The geometry of the syn-rift reflectors further suggests the reactivation of this fault during the Miocene. This subvertical WNW-striking structure is interpreted to be analogous to the Telegdi-Roth Line as a dextral wrench fault, but with a smaller offset which could not be determined due to the poor seismic coverage. The compressional component along this reverse fault apparently increases to the WNW and the fault itself
terminates before the Ikervár high, again, by detachment on a basement thrust plane.

5.2.9 AREA I (IKERVÁR)

Area I represents a transitional area between the western part of the basin, characterized by a large amount of extension along detachment faults, and the eastern part of the basin, where the Eoalpine thrust contacts in the basement are relatively intact. Three dip-oriented seismic illustrations are reproduced in Panel 13, since no reasonable strike section was found in this area.

The relatively thick Neogene sequence in section I1 indicates the gradual deepening of area I to the SW to the Zala Basin, as opposed to the relatively shallow basement shown in sections I13 and I16. These latter sections cover the transitional area between the Danube Basin and the Zala Basin (cf. Figs. 5.3 and 5.4a).

In the pre-Senonian basement, the northwestern end of section I1 is fairly similar to the previously discussed I5 section in Fig. 3.7. The top of the Penninic can be easily picked since it separates strong SE-dipping reflectors from the less coherent, block-faulted basement from above. Here, similarly to the nearby outcrops of Eisenberg and Hannersdorf (Figs. 5.3 and 5.4d) the low-grade Paleozoics of the Upper Austroalpine directly overlie the Penninic greenschists, pointing to a major tectonostratigraphic omission (up to 11 km, see subchapter 5.4.1.3.2 for details).

The hanging wall of this detachment fault is dissected by a number of more steeply dipping normal faults. Miocene extensional reactivation of a subhorizontal Eoalpine thrust fault can be observed at the eastern end of section I1. This structure becomes very prominent farther to the S, in the neighboring area A (Fig. 5.4d).

In sections I13 and I16 several intra-basement thrust units were interpreted and calibrated along strike by the wells on the Ikervár high (Ike-6 and Sót-2 wells, Paleozoic metamorphics) and by the Sót-1 and Kám-1 wells (Hauptdolomite). As to the Rába "Line"
problem, two points should be emphasized in area I. Firstly, the Rába Line of many (e.g. Wein, 1975; Kőrössy, 1987) runs between wells Sőt-2 and Sőt-1, but the coherently SE-dipping intra-basement reflectors exclude the presence of a subvertical fault of any type (sections I13 and I16). Instead, this contact is an Eoalpine thrust plane. Secondly, the Rába fault as it is defined in this work developed from the overthrust plane shown in the northwestern flank of the Ikervár high by its increasing extensional reactivation along strike, to the NE (Fig. 5.4d).

5.2.10 AREA A (ANDRÁSHIDA)

Because of the major change in the strike of pre-Senonian basement structures (Fig. 5.4d), the distinction between strike and dip sections becomes somewhat arbitrary. At any rate, two NW-trending and four NE-trending seismic sections were reproduced in Panels 14 and 15.

The geometry interpreted in sections A6 and A8 is similar to that of section I1 in the previous I area. In section A6 the top of Penninic can still be picked based on several strong SE-dipping reflectors. Note that the Sarmatian strata at the lowermost part of the post-rift succession are fairly thick (400-500m) in the center of the basin but gradually pinch out towards the basin margin.

A SW-oriented progradation within the Pannonian sequence is best expressed on NE-trending profiles (e.g. sections A16 and A17), in contrast to the above described areas.

Also on sections A16 and A17 a deep and about 7 km wide, very prominent syn-rift trough shows up. This slightly asymmetric trough is bounded by steeply dipping normal faults on both flanks, the master fault being on the southwestern flank. Moreover, a number of smaller half-grabens were interpreted to the SW of this major trough. These small features are also dominantly controlled by NE-dipping normal faults, but these NW-
trending faults (Fig. 5.4d) are apparently detached on a subhorizontal surface between about 2-3 km beneath the top of the pre-Tertiary basement. Sections A6 and A8 suggest that this surface is in fact an Eoalpine overthrust plane.

Aside from the deep Miocene trough, only one structural feature seems to crosscut this décollement surface. This WNW-trending structure starting from the SE displays the geometry of a subvertical reverse fault on section A20, a normal fault on section A17, an inverted half-graben in section A16 and a local thrust in section A15 (Fig. 5.4d). These sudden along-strike structural changes indicate the strike-slip origin of this fault, similarly to that of the Telegdi-Roth Line (see area D).

Furthermore, since a thick Neocomian marly sequence (Sümeg Marl) is present in section A20 according to the nearby Nagytílaj(Nt)-2 well (Körössy, 1987; Kázmér, 1987b; Fig. 5.4c), the age of movement along this fault is certainly post-Barremian. The overlying thin Senonian onlaps the eroded surface of the Neocomian, post-dating the faulting. These Senonian strata, however, are also erosionally truncated by the syn-rift sequence in a geometry clearly suggesting the reactivation of this fault during the Middle Miocene.

The most striking Eoalpine structural feature is revealed by the map view of intra-basement thrust units (Fig. 5.4d). In sections A6 and A8 these features can be identified as the southwesterly continuation of the same thrusts described in areas I, K and D. These thrusts, however, do not maintain this orientation, but instead they display a pronounced curvature to the SE. Therefore the same structural trend can be followed on sections A15, A16 and A17. One of the thrust surfaces was apparently reactivated during the Middle Miocene syn-rift phase and shows an approximate 2 km maximum horizontal offset at the apex of the curve. This geometry implies a northeastward movement of the hangingwall during extension along a subhorizontal detachment surface. Since the hangingwall block exclusively consists of Hauptdolomite according to drilling evidence (Ger-1, Va-1 and Nt-4 wells; Fig. 5.4d), the detachment is placed in the underlying Veszprém Marl (Fig. 5.7).
5.2.11 AREA S (SZENTGOTTHÁRD)

Similarly to areas N and V the seismic reflection data in the vicinity of Szentgotthárd are represented mostly by sections acquired during the seventies. Therefore I did not attempt to illustrate these lines on a panel.

One published seismic section (S7 section on Fig. 3.5), however, clearly shows the main structural character of this area. The down-to-the-SE low-angle normal fault bounding the South Burgenland basement high may be the southwestward structural continuation of the Rába fault s.s.(cf. Rumpler and Horváth, 1988), but it is to be noted that they are separated by the Ikervár area displaying no normal offset (Fig. 5.4d).

Two nearby wells (Szg-1,-2; Fig. 5.4c) drilled very low-grade to low-grade Upper Austroalpine metamorphics on top of the South Burgenland basement swell. Similarly to area I, however, the dip sections usually reveal a subtle change in seismic reflectivity. Strong reflectors between about 2 and 3 s TWT time (~5-7 km) on section S7 (Fig. 3.5) probably emanate from Penninic metamorphites. This interpretation is supported by the outcropping contact between the Upper Austroalpine and the Penninic in the Eisenberg Mts., some 30 km to the N and by the nearby Güssing-1 well in Austria. This later well drilled into Penninics under thin Pannonian sedimentary cover on the crest of the South Burgenland basement high (Fig. 5.4d).

5.2.12 AREA Z (ZALATÁRNOK)

Unfortunately subarea Z is covered by relatively short (10-20 km) seismic reflection sections of relatively poor quality gathered during the seventies (Fig. 5.4a). Therefore no seismic illustration is reproduced on panels from this subarea.

The structure of area Z, however, is best illustrated by an E-trending composite of sections Z20 (reprocessed) and Z21 (Plate 9). This line seems to be a dip line both for the Neoalpine and the Eoalpine structures across the deepest part of the Zala Basin. The
Neoalpine structure is dominated by a major low-angle normal fault similar to areas S, A and I. Based on borehole evidence the hangingwall of this fault consists of Upper Austroalpine Triassic to Senonian carbonates correlative with their outcropping counterparts in the Bakony Mts (Fig. 5.4c). The footwall with its distinctive seismic facies (strong, E-dipping reflections) was not reached by drilling in Hungarian territory but it was penetrated some 10 km to the W in Slovenia, according to Kröll et al. (1988). The medium-grade crystalline rocks found in the Dankovci-1 well (Fig. 5.4c) belong to the Middle Austroalpine which outcrops farther to the W in the Pohorje Mts (see Fig. 5.1). The fault probably trends to the NW and I term it the Mura fault (after River Mura in Slovenia).

The hangingwall displays internal structures revealed by strong reflector packages dipping to the E. These features are Eoalpine in age since the Senonian strata seal them. Although in this area their 3D geometry could not be mapped in detail, I interpret them as overthrust planes analogous to those described in areas D, K and A (Fig. 5.4d). Wells nearby with older-on-younger Mesozoic rocks provide evidence for these thrust contacts (e.g. Bm-I, Zeb-2; see Fig. 5.4c). The eastern part of area Z is dominated by the E-W trending, poorly understood Eocene Zalatárnok-Bak graben (see Z11 section on Fig. 3.8).

5.3 REGIONAL STRUCTURE TRANSECT BASED ON DEPTH-CONVERTED REFLECTION SEISMIC SECTIONS

The location of the 1:200,000 scale regional section (Plate 10) is shown in Fig. 5.1b. The section starts in the N on the European foreland and terminates in the S to the S of Lake Balaton, at the Mid-Hungarian shear zone.

In the N, the structure section is based on a section published by Wessely (1983, 1988). Structural units in this region can be subdivided into three major groups (or levels, see Wessely, 1988). The lowermost level is the crystalline mass of the European foreland
which dips gently to the S. In the Vienna Basin the autochthonous Mesozoic cover is usually preserved. This cover, however, is missing in the crest of the Bohemian basement promontory (Wessely, 1987). On top of the European foreland a thin succession of flysch and molasse overlying the Mesozoic can be found as far as 35 km to the SE from the leading edge of the Alpine thrust belt (see Berndorf-1 well).

The next level is the allochthonous Alpine nappe complex. Its well-known upper part outcrops to the W of Vienna (Fig. 5.1). It belongs to the Upper Austroalpine system, which in the Northern Calcareous Alps can be subdivided into three main units (e.g. Hamilton et al., 1990). These are from top to bottom: Upper Limestone Alps and Graywacke zone (Juvavicum), Ötscher nappes (Tirolicum, recently called Göller nappe system) and Frankenfels-Lunz nappe system (Bajuvaricum). All these nappes are "cover nappes" (e.g. Tollmann, 1987c), i.e. they were detached from their metamorphic basement. These nappes exclusively consist of unmetamorphosed Mesozoic sequences, except for the uppermost Juvavic nappe which has a low-grade metamorphosed Paleozoic substratum (Graywacke zone). All these units are structurally underlain by Middle and Lower Austroalpine thrust sheets which also outcrop along strike. The presence of the Penninic unit at depth is problematic in the area of the Vienna Basin due to lack of well control (Wessely, 1988).

The uppermost level is represented by the Neogene succession of the Vienna Basin. The structure section crosses the southwestern corner of this basin (Fig. 5.1), where normal faults bound the 2-3 km deep Neogene basin. These normal faults are shown to sole out and merge with the base of the underlying Alpine nappe complex in Plate 10. Note that this is the only modification I made on the original sections of Wessely (1988), who thought that the normal faults also cut and offset the European foreland. These normal faults accommodated sinistral strike-slip movements required by many authors (Royden et al., 1982; Fodor et al., 1989; Fodor, in press) for the opening of the Vienna
pull-apart basin. The inferred left-lateral offsets along these major faults, however, could not be documented (Wessely, 1988).

The Vienna Basin is separated from the Danube Basin (or Little Hungarian Plain) by a basement high trending perpendicular to the section. The basement here consists of Lower Austroalpine units outcropping in the nearby Leitha Mts. (see Fig. 5.3). Farther to the S the smaller Mattersburg Basin appears to be bounded by a S-dipping normal fault.

The section crosses the Austrian/Hungarian border just to the N of Sopron and from there follows the trace of regional seismic section D (Panel 1) described in Chapter 3. Thus between the Sopron area and the Dabrony-1 well the section is based on the MK-1 crustal section (Plate 3) of Ádám et al. (1984). Farther to the S the section follows the non-crustal continuation of the MK-1 line (Plate 4) through the Bakony Mts. Note that this part has been processed only to 4 s TWT time. This part of the section was also constrained by the surface geology derived from a number of maps listed in Appendix C.

In the northwestern part of the Danube Basin the pre-Tertiary basement exhibits a characteristic basin-and-range morphology. Subbasins (Mattersburg, Nagycenk, Csapod, Kenyeri) are separated by basement highs (Leitha, Sopron, Pinnye, Mihályi). All of these subbasins are controlled by major SE-dipping Middle Miocene normal faults (cf. earlier interpretations with vertical faults, e.g. Körössy, 1981, see Fig. 4.12). The crustal section (Plate 3) clearly shows that at least two of these faults (Fertő and Répce) maintain their low-angle dip (~30-40°) to middle crustal depth. Moreover, the Fertő fault appears to merge into a prominent surface which I consider the base of the Austroalpine nappe complex. The midcrustal geometry of the Rába fault is not clear because the data deteriorate in the southeastern part of the crustal profile.

The average dip (~5°) of the European foreland at the northwestern end of the section is well constrained by the Raipoltenbach-1 and Berndorf-1 wells. Extrapolating this dip to the SE, the top of the European foreland can be tied into the northwestern end.
of the crustal section (Plate 3). A prominent reflection doublet at 11 km depth is
interpreted to originate from the boundary of the European plate and the overriding Alpine
edifice. This reflection event can be correlated farther to the SE with a slightly higher dip
(−10-15°), to about 20 km depth. Poor seismic data quality does not permit to follow this
surface farther to the SE of the Mihályi high (Plate 3). Subhorizontal to slightly SE-
dipping strong reflector packages below this surface are considered to emanate from
crystalline rocks of the European Bohemian Massif. Beneath the Pinnye High at about 14-
18 km some strong NW-dipping reflectors are tentatively interpreted as related to a 10-15
km wide Mesozoic half-graben on the distal edge of the European passive margin. The
boundary fault was dipping toward the Jurassic Penninic ocean, i.e. to the S.

Regarding the depth of the Mohorovicic discontinuity along the section, the map of
Posgay et al. (1991) shows this surface in an elevated position at 26 km beneath the
Mihályi high. In the central part of the crustal seismic section (Plate 3) very low frequency
reflectors between 9-10 s TWT may correspond to the Moho. Below the Eastern Alps the
depth of the Moho is not known (Posgay et al., 1991).

In the southeastern part of the Danube Basin the pre-Tertiary basement displays a
general monoclinal dip to the NW. Here the Eoalpine structures are reasonably well-
known near the top of the pre-Senonian basement, based on the interpretation of the
industry seismic profiles described earlier. There are a number of NW-verging nappe
structures that display an Eoalpine deformational style similar to their counterpart Upper
Austroalpine nappes of the Eastern Alps. Farther to the SE there is only one seismic
profile crossing the Bakony Mts, shown on Plate 4. Although the quality of this section is
relatively poor, some wells and pre-Tertiary basement outcrops in the SE aided the
interpretation.

There are two synclines (Devecser and Halimba) superimposed on the overall
synclinal shape of the Bakony Mts. In map view (Fig. 5.4d) these synclines are flanked by
anomalously wide (10-15 km) Hauptdolomite successions. Allowing for an average 30°
dip of the Hauptdolomite as reported from both outcrops and wells (e.g. Bencze et al.,
1990; Körössy, 1987), this would mean a 5 km thick sequence. Since the maximum
observed thickness of the Hauptdolomite is 1500 m (Fig. 5.7), I postulate its internal
repetition by thrusting. The poor seismic data unfortunately do not resolve any
interpretation. The backthrusting at the southern edge of the Devecser syncline over the
northern margin of the Halimba syncline is documented in outcrops (see e.g. map of
Császár et al., 1978). The detachment of the Hauptdolomite from its base occurred along
the Carnian Veszprém Marl (Fig. 5.7), which might have accumulated in a triangle shape
subsurface area between the two synclines.

The two synclines appear to float above two regional thrust surfaces correlatable
from their outcrops in the Balaton Highland to their subcrop in the Danube Basin.
Whereas the structurally higher Veszprém thrust is associated with the Carnian
detachment level, the deeper Litér thrust generally follows a Middle Triassic detachment
level. These thrusts do not have a clear seismic expression in the seismic section of Plate 4
and therefore the presented interpretation is not more than a tentative first attempt. The
good conductivity zone determined by magnetotellurics is also shown in Plate 4 and it
shows a loose correlation with the above described detachment levels.

To constrain the deep structure of the transect across the Bakony Mts., crustal line
MK-3 shown in Plate 6 was projected into the regional structure section from some 60 km
to the NE (see Fig. 3.1). The upper crustal part of this line displays the same overall
geometry as observed on line MK-1 (Plate 4), which justifies the long-distance along-
strike projection. The middle and lower crustal part of this crustal line shows a major
anticlinal feature beneath the Csatka Basin. The corresponding thrust seems to root into a
southeasterly dipping surface marked by a prominent reflector package. This surface is
inferred to be the base of the Upper Austroalpine system, based on the downdip
extrapolation of this contact from the crustal section shown in Plate 3. The overall synclinal geometry of the Bakony Mts. might be caused by two NW-verging deep thrusts involving the crystalline basement of the Middle and Lower Austroalpine. Probably these deep thrusts are responsible for the along-strike appearance of the Velence pericline shown in Fig. 4.13.

In the transect the correlative anticline under the Lake Balaton is more distorted because of the younger, Early Miocene Balaton "Line". As discussed in detail in Chapter 3, this fault does not have a clear seismic expression. Its reverse fault character at shallow depth was documented by drilling (e.g. Balla et al., 1987; Körtössy, 1990). In the transect the Balaton fault was placed immediately to the N of the Karád wells where Dinaric-type Paleozoic carbonates were found (Bérczi-Makk, 1988a). At depth I postulate the flattening of the Balaton Line at the base of the Austroalpine nappe system. Since the large-scale Early Miocene right-lateral movements along this fault occurred synchronously with the last thrusting of the Alpine units over the European foreland, the Balaton Line is essentially a huge tear fault. Thus, it should be detached at the base of the eastward escaping unit shown in the transect. The detachment surface is likely to be located within the Penninic unit (see Chapter 8 for a discussion).

As an indirect argument for the flattening of the Balaton Line at lower crustal depth the results of Heitzmann (1987) and Heitzmann et al. (1992) are cited here. These authors were able to demonstrate the mid-crustal northward flattening of the Insubric Line which appears to be the westernmost extension of the Periadriatic-Balaton Line system.

To sum up, perhaps the most striking result of the regional transect is that the shallowly dipping (5°-10°) European foreland indeed projects below the Alps and the Pannonian Basin and can be traced some 200 km(!) to the SE. Thus the transect seems to fully support the view of Wessely (1987) on the subsurface continuity of the Bohemian basement spur. Note, however, that this finding is at odds with the interpretation of
Tomek and Hall (1993), who proposed the abrupt deepening and steep dip (70°-80°) of the European plate along strike, about 100 km to the NE, in the Slovakian sector of the Vienna Basin.

5.4 ALPINE TECTONICS OF THE NW PANNONIAN BASIN

The summary on Alpine evolution is based mainly on the preceding seismic interpretation. This subchapter begins with the present-day situation and goes backwards in time. The early Alpine history of the area will not be discussed in any detail, because the seismic profiles do not resolve any structure related to this period.

5.4.1 NEOALPINE EVOLUTION OF THE STUDY AREA

I subdivided the Neoalpine structural stage into four periods, i.e. Quaternary-Recent neotectonics, Late Miocene-Pliocene post-rift tectonics, Middle Miocene syn-rift tectonics and the Early Miocene "escape" tectonics (see Fig. 5.2). As to the syn-rift/post-rift boundary I place it stratigraphically earlier than others (e.g. Royden et al., 1983b). Commonly this boundary is placed at the Pannonian/Sarmatian boundary, but in my view the boundary should be placed between the Upper and Middle Badenian, some 3-4 Ma earlier (Fig. 3.4, see also Appendix B for a discussion of the time scale).

Widespread Late Badenian/Sarmatian strike-slip faulting and the small-scale inversions of the Transdanubian Central Range are described as part of the post-rift succession, although these faults are commonly cited as the evidence for syn-rift strike-slip faulting (e.g. Royden, 1988).

Another new element of this summary is that the syn-rift period is further subdivided into an Early-Middle Badenian "wide-rift" style and a Karpatian "metamorphic core complex" style extension, following the terminology of Buck (1992).

Early Miocene (Late Egerian-Eggenburgian-Ottnangian) escape tectonics are poorly
defined but follow the formation of the underlying succession of the Paleogene Basin. Note that escape tectonics are limited by many to the Paleogene (e.g. Kázmér and Kovács, 1985; Fodor et al., 1992).

5.4.1.1 QUATERNARY-RECENT TECTONICS

The most spectacular evidence for the gentle but widespread Quaternary compression (Fig. 5.2) in the NW Pannonian Basin comes from reflection seismic data. This industry seismic profile shown in Fig. 5.8 is two times vertically exaggerated and extends almost to the deepest part of the Danube Basin. The prominent syncline involving the Pannonian section appears to be caused by compression. Compaction as the dominant cause for the synclinal shape can be excluded since even the uppermost Pannonian reflectors display a near-parallel geometry. Unfortunately the uppermost 400 ms of the section (~400 m) was muted, but from the map of Frányó (1971) shown in Fig. 5.9 it is clear that the thickness of the Quaternary mimics the underlying syncline. The present-day elevation of Danube terraces was mentioned in Chapter 4 (Fig. 4.21) as independent evidence for the Quaternary subsidence in the Győr subbasin.

As the center of the syncline deepened, the flank of the Danube Basin was uplifted and a significant part of the Pannonian sequence was removed by erosion (Fig. 5.9). This erosion was documented in the monography of Jámbor (1980). The amount of missing Pannonian sequence is best determined in the southwestern part of the Bakony Mts., where a number of Pliocene volcanoes became buttes following the Quaternary uplift (Fig. 5.9). Jámbor (1980) measured the thickness of the Pannonian preserved beneath the resistant basalt cap, which provides a minimum estimate for the eroded succession.

The observations on the Quaternary subsidence (basin center) and uplift (basin flank) of the NW Pannonian Basin are incorporated into an idealized cross-section in Fig. 5.10., which also shows the geometry of the underlying Tertiary basin complex.
Fig. 5.8. Two times vertically exaggerated industry seismic profile (G7 section) showing inversion tectonics.
Fig. 5.9. Basin inversion associated with Quaternary compression in the study area, compiled from Frányó (1971) and Jámbor (1980).
Fig. 5.10. Schematic structure-section across NW Hungary. Approximate trace of section is shown in Fig. 5.1.
Interestingly enough many argued that during the Late Pliocene(?) the original course of the Danube differed significantly from its present-day course (see Frányó, 1971; cum. lit.). During that time the Danube may have flowed south from the Vienna Basin through the Zala Basin and directly to the Dráva Basin (Fig. 5.1). The shift to its present-day course might be attributed to the Pleistocene uplift of the southwestern continuation of the Transdanubian Central Range forming an effective river divide.

Independent evidence for the Quaternary inversion in the NW Pannonian Basin comes from the present-day stress field. Based on recently performed in-situ and borehole breakout stress measurements, the Danube Basin is in compression with the maximum horizontal stress oriented NW-SE (A. Becker, pers. comm., see also Dövényi and Horváth, 1991; Gerner, 1992).

The above described structural inversion can be understood in terms of late stage compression (Kooi and Cloetingh, 1989; Kooi, 1991) often associated with older rift basins (e.g. North Sea, Cloetingh and Kooi, 1992).

5.4.1.2 LATE MIOCENE-PLIOCENE POST-RIFT TECTONICS

Seismic data of the Danube Basin show that Pliocene basalt magmatism (Tapolca Basalt) is more widespread in the subsurface than previously inferred solely from the distribution of outcropping volcanic centers (e.g. Figs. 5.1 and 5.9). Since this late volcanism follows the more localized Middle Miocene intermediate volcanism with a well defined time gap of about 4-5 Ma, their genetic connection seems to be far from obvious, and in my opinion remains an open problem (cf. Embey-Isztin et al., 1990; Szabó et al., 1992). The same holds for the Styrian Basin as well (cf. Kurat et al., 1980; Ebner and Sachsenshofer, 1991).

The huge Badenian(?)-Early Pannonian Pásztori Trachyte volcano, however, in the middle of the Hungarian Danube Basin (Fig. 5.4d) seems to record the initial stage of
continental breakup beginning with narrow rifting (sensu Buck, 1990). Following a stage of wide rifting (see subchapter 5.4.1.3.1) when extension was more regionally distributed over a number of detachment faults across the basin (see below), the Pásztori volcanism occurred in local extensional zones. I speculate that this site in the center of the basin heralds the concentration of continental extension into one single rift graben (Fig. 5.12). This rift failed, however, because back-arc extension ceased soon in the NW Pannonian Basin due to the geometric boundary conditions imposed by the thick crust of the European foreland (Fig. 5.1).

An undrilled and therefore poorly understood magnetic anomaly in the Slovakian side of the Danube Basin near Kolárovo (Fusán et al., 1987) may well be associated with a similar-size paleovolcano. Indeed the shape and size of the Kolárovo magnetic anomaly is very similar to that of the Pásztori volcano.

Late Miocene (Sarmatian to Late Badenian) strike-slip faulting in the Bakony Mts. (e.g. Mészáros, 1983) has been confirmed by seismic reflection data presented in this thesis. There are more strike-slip zones than previously thought, even though the offsets cannot be determined (Fig. 5.3d). These fault zones are not exclusively Miocene in age (cf. Kőkay, 1976 1985), but represent reactivated Cretaceous faults (see below), in agreement with the original proposition of Telegdi-Roth (1935) and the theoretical speculation of Tari (1991).

It is to be emphasized that the frequently observed Sarmatian transpression associated with these fault zones amounts to a structural inversion. Microtectonic studies of these faults indeed indicate a N-S directed compressional paleostress field (Csontos et al., 1991). Therefore these strike-slip faults are better interpreted in terms of a compressional stage taking place shortly after syn-rift extension, rather than syn-rift wrenching (cf. Royden, 1988 and many others). Transpression clearly occurred during the post-rift period and there is evidence for this regional inversion in other regions of the
Pannonian Basin as well (e.g. Tari, 1992c).

5.4.1.3 MIDDLE MIOCENE EXTENSIONAL TECTONICS

During the Middle Miocene syn-rift phase two extensional systems are distinguished (see Fig. 5.2). The earlier system accommodated extension in a ENE-WSW direction, whereas the subsequent, almost perpendicular extensional system had an NW-SE orientation. This relationship was first observed by Ratschbacher et al. (1990) based on microtectonic measurements in the Rechnitz Window group. These authors observed an Oligocene to Early Miocene (?) ductile extensional event (D2) trending ENE. This extension apparently occurred at midcrustal depth (~10 km, Koller, 1985). Ratschbacher et al. (1990) also reported a Middle to Late Miocene brittle E-W extensional event (D4).

These observations can be reconciled with the results of seismic interpretation. The earlier ductile extensional event corresponds to the formation of the major Rechnitz detachment fault and it is manifested by the Rechnitz-Wechsel metamorphic core complex system (~70-90 km ENE-WSW extension). During the subsequent extensional period (~30-50 km NW-SE extension) a series of smaller, brittle detachment faults formed (Fertő, Ikva, Répce and Rába) in a more distributed deformational style. Obviously, these extensional modes overlapped in space and time in certain parts of the NW Pannonian Basin.

Thus following the terminology of Buck (1990), for the NW Pannonian Basin and the adjacent part of the Eastern Alps an earlier metamorphic core complex and a subsequent wide rift style of Middle Miocene upper crustal extension can be distinguished.

The dominant mode of continental extension is a function of the original heat flow and crustal thickness (Buck, 1990; see Fig. 5.11). It is hard to evaluate the heat flow at the onset of extension but it was at least as high as the recently observed maximum values in the NW Pannonian Basin (90-95 mW/m²; Dövényi and Horváth, 1988). As to crustal
thickness the only slightly extended Bakony Mts to the E and the unextended central zones of the Eastern Alps to the W may show the paleothickness (32-45 km) of the Moho just prior to the Miocene extension (e.g. Posgay et al., 1991). At present the Mohorovicic discontinuity is at 26-30 km depth beneath the Styrian and Danube Basins and the average heat flow values scatter between 60-70 mW/m². Therefore based on the results of Buck (1990) a temporal progression of extensional style is proposed here: a Karpatian core complex style extension was followed by Badenian wide-rift style extension (Fig. 5.11). In the center of the Danube Basin narrow-rift style extension might have began during the Sarmatian and the Lowermost Pannonian, but this extension could not develop into a well-defined rift-graben.

5.4.1.3.1 WIDE RIFT STYLE OF SYN-RIFT TECTONICS

In the northwestern part of the Hungarian Danube Basin three major detachment faults can be mapped based on seismic data - structurally from deepest to shallowest: the Ikva fault, the Répce and the Rába fault (Fig. 5.4d). These generally NE-trending listric faults have a variable normal offset along strike, ranging from a few kilometers to about 10 km. The displacement direction is invariably down-to-the-SE. The "enigmatic" Rába fault is just one of these low-angle normal faults and it has very little, if any, strike-slip offset. The magnetotelluric Rába Line of some (e.g. Balla et al., 1990) does not coincide with the Rába fault as it is defined in this work. As will be shown later, the magnetotelluric data can be used to follow the subsurface extent of a Cretaceous overthrust contact.

The low-angle normal faults created considerable relief (up to 3 km) and a basin-and-range topography in the northwestern part of the basin (Panels 4-7). The asymmetry of the basement highs clearly controlled the distribution of alluvial fans. The entry points of footwall-sourced clastic fans seem to be directly controlled by along-strike variations of normal offset, as shown in Fig. 5.12 for the Répce fault.
MODES OF CONTINENTAL EXTENSION

NARROW RIFT

NW PANNONIAN BASIN

WIDE RIFT

CORE COMPLEX

Fig. 5.11. Modes of continental extension tectonics from Buck (1990) and the classification of the NW Pannonian Basin.
RÉPCE DETACHMENT FAULT, CSAPOD SUBBASIN, NW PANNONIAN BASIN, HUNGARY

Fig. 5.12. Schematic 3D diagram of the Répce detachment fault. View is to the N.
Earlier I addressed the problem of the widely held assumption that Cretaceous overthrust planes were reactivated during the Miocene as low-angle normal faults in the Pannonian Basin (e.g. Grow et al., 1989; Tari et al., 1992). Based on the seismic illustrations shown in Panels 4-15 it is now clear that Neoalpine low-angle normal faults indeed interacted with abandoned Eoalpine thrust fault planes. Such interaction, however, seems to be more complex than was expected.

One can outline three possible means of interaction between normal faults and pre-existing thrust faults as follows. (1) One is that an earlier thrust plane is extensionally reactivated along its entire length (e.g. Ratcliffe et al., 1986). (2) Another scenario involves listric normal faulting where the relatively steep upper part of a normal fault flattens out at depth, at a regional décollement level (e.g. Bally et al., 1966). (3) Yet another geometry should be taken into account where shallow thrusts are cut by extremely low-angle normal faults (e.g. Wernicke et al., 1985). Recently Ivins et al. (1990) reviewed the factors, such as geometry, intact/pre-existing fault strength and fluid pressures, which determine whether extensional reactivation will or will not occur, although they did not specifically study the above described geometries.

In the NW Pannonian Basin Middle Miocene, reactivation of Middle Cretaceous thrust planes by low-angle normal faults occurred dominantly in the second manner (see area C, Panels 4-7 and area M, Panel 8) and less typically in the first manner (see area D, Panel 10 and area A, Panels 14-15). The geometry where shallow thrusts are cut by extremely low-angle normal faults was not observed.

The conclusion is that even though reactivation occasionally occurred, Cretaceous overthrust and Miocene detachment fault planes rarely coincide at the top of the basement or at shallow depth beneath it. That is why in map view the Répce and Rába faults appear to overall "ignore" the Austroalpine nappe contacts (Fig. 5.4d). Towards deeper intra-basement levels, however, they frequently merge into each other, as illustrated by the
cross-sectional geometry shown on Panels 4-8.

The *relative* timing of the Ikva, Répce and Rába detachment faults shows a progression to the SE as inferred from cross-cutting relations. More specifically, detachment faulting tends to propagate into the hangingwall, into the direction of transport. This deformational sequence is regarded as characteristic for low-angle normal fault systems, according to John (1987).

The southeastern side of the Danube Basin displays a markedly different style and polarity of Miocene extension. To the SE of the Rába fault, its hangingwall is characterized by NW trending, steeply dipping Miocene normal faults forming a series of high blocks and basins in between (Fig. 5.4d). Such a general geometric pattern was already inferred from potential field data (e.g. Körössy, 1987). Similarly to the above described post-rift strike-slip faults, these steep normal faults cannot be traced across the Rába fault and thus are limited to the hanging wall of the Rába fault, in some cases soling into a relatively shallow detachment surface (Panels 14 and 15).

5.4.1.3.2 METAMORPHIC CORE COMPLEX STYLE SYN-RIFT TECTONICS

Core complex type extension in the transition zone between the Eastern Alps and the Pannonian Basin was first recognized by Tari and Bally (1990). They compared the formation of the Rechnitz Window group to the well-known metamorphic core complexes of the western US (Davis and Lister, 1988; Lister and Davis, 1989; cum. lit.). In the following I summarize the evidence supporting this interpretation.

The seismic sections of Panels 4-7 show that the clearly defined Hungarian east plunge of the Rechnitz window displays a dome-like feature and that its tectonic contact with the overlying Austroalpine units is a major detachment fault. This observation leads to asking what is the specific nature of the upper and lower plate of this system, where is the breakaway of the core complex-forming detachment fault and what is the direction and
magnitude of extension?

Figs. 5.13a,b show the transitional zone between the Eastern Alps and the Pannonian Basin. Due to the large-scale continental extension described below I introduce the term Rába River extensional corridor for this region. It is important to realize that there are tectonic contacts in this area where a significant part (up to 10 km) of the Austroalpine system is missing in contrast to neighbouring areas where it is complete (Fig. 5.14). These major omission contacts are as follow (see Fig. 5.13c for location).

1) Anger area, where the Anger Crystalline (Upper Austroalpine) lies right on top of the Strallegg Gneiss (Lower Austroalpine), according to Flügel and Neubauer (1984), and Neubauer et al. (1992a).

2) Wechsel area, at the northeastern margin of the Wechsel Window, where the deepest Wechsel unit within the Lower Austroalpine has a fault contact with the "Grobgneiss" complex (Eselsberg Gneiss, Neubauer et al., 1992a; Neubauer and Frisch, in press).

3) Kirschlag area, at the northeastern margin of the Berstein Window (see also Fig. 5.15) where the Middle Austroalpine is right on top of Penninic greenschists (Pahr, 1980; 1984).

4) Eisenberg area, where the largest omission of tectonostratigraphy is suggested by the outcrop of the Upper Austroalpine Hannersdorf Paleozoic having a fault contact with Penninic greenschists (Tollmann, 1977; Schmidt et al., 1984). Note that a similar omission was documented in the subsurface to the E of the Rechnitz Window (see subchapter 5.2.1, Panels 4-7).

Note that along strike, in the Pohorje Mts. (Fig. 5.13c) the Middle Austroalpine is at least 5 km thick (Hinterlechner-Ravnik and Moine, 1977) and in the Little Plain area of the Danube Basin the Middle Austroalpine (?) and the Lower Austroalpine together is at least 10 km thick (see section C7 on Panel 7). These supposedly intact tectonostratigraphic columns provide the basis to estimate the thickness of the missing section (Fig. 5.14).
Fig. 5.13a. Surface geology of the Rába River extensional corridor of Austria/Hungary. Sources listed on map.
Fig. 5.13c. Index map of the Rába River extensional corridor Austria/Hungary. See text for detailed explanation.
Fig. 5.14. Major omissions in tectonostratigraphy in the Rába River extensional corridor. Lower part shows the estimated thickness of missing section. Location of these columns is shown in Fig. 13c. Based on various sources listed in the text.
Similar omission contacts of lesser magnitude can be observed on the detailed tectonic map of the Rechnitz Window group (Fig. 5.15). Supporting subsurface evidence comes from the Maltern-1 well drilled at the southern margin of the Bernstein Window (Fig. 15b). In this well beneath 55 m of Lower Austroalpine Grobgneiss the entire Wechsel succession was represented by 45 m thick (!) albitegneiss on top of Penninic serpentinite (Pahr, 1977).

These prominent omission contacts can only be explained in terms of large-scale detachment faulting involving the whole transitional zone between the Eastern Alps and the Pannonian Basin. I propose that the major omission contacts shown in Fig. 5.14 are related to the very same detachment fault (top to the NE-ENE) that formed the Rechnitz metamorphic core complex (Fig. 5.16).

There are a number of original elements on the cross-section shown in Fig. 5.16, which is drawn parallel with the dominant displacement direction (WSW-ENE) during the core complex-type extension. The upper crustal part of the cross-section displays the main geometric characteristics of a core-complex forming detachment fault, i.e. its low-angle character and the updoming of its middle portion (Spencer, 1984; Wernicke, 1985). Around and on top of the Rechnitz core complex the upper plate is extremely thinned by a number of internal low-angle normal faults which developed within the hangingwall of the large Rechnitz detachment fault. In this way the Lower Austroalpine units thinned sufficiently to expose the lower plate. Note that to the W of the Rechnitz window the detachment fault climbs up-section with respect to the footwall (or lower plate), even though it dips westward. The westward dip of the Penninic/Austroalpine boundary immediately to the W of the Rechnitz window group was estimated as about 10° by Walach (1977) and Walach and Wöber (1987) based on magnetic anomalies. The only possible candidate for the upper/lower plate boundary to the W of the lower plate culmination is at the eastern edge of the Graz Paleozoic (Fig. 5.13c), where the whole
Fig. 5.15a. Tectonic map of the Rechnitz metamorphic core complex (simplified after Pahr, 1984; Hermann and Pahr, 1988).
Fig. 5.15b. Index map of the Rechnitz metamorphic core complex.
Fig. 5.16. Conceptual crustal dip section across the Rába River extensional corridor. Vertical exaggeration is about twofold.
Middle Austroalpine is missing. The Graz Paleozoic itself is shown as a major extensional allochthon, where the extension has not been sufficient to expose the lower plate (Fig. 5.16).

Therefore the breakaway of the Rechnitz detachment fault should be placed to the W of the Graz Paleozoic. The contact between the Graz Paleozoic (upper plate) at its western edge and the underlying Middle Austroalpine (lower plate) is indeed a documented down-to-the-E low-angle normal fault (Krohe, 1987; Fritz et al., 1991), for which most authors postulated a Cretaceous age. The original position of the headwall breakaway was certainly to the W of its present-day erosional exposure at the western end of the Graz Paleozoic or in the Pohorje Mts. (Fig. 5.13c). An educated guess would place the inferred pre-erosional position of the breakaway along the crest of the Koralpe, where the present-day morphology seems to reflect its breakaway range character (i.e. upward flexure toward the eastern side of the range, cf. Wernicke, 1985).

As to the map view, the upper/lower plate contact can be traced with reasonable confidence based on stratigraphic omissions (Fig. 5.13c). To the S of the Rechnitz window, in the Eisenberg Mts. the Upper Austroalpine Hannersdorf and Sulz Paleozoic (Fig. 5.13c) is obviously in an upper plate position separated from the underlying lower plate Penninics by the Rechnitz detachment fault. To the NE of the Rechnitz Window group the upper/lower plate contact can be placed on the surface at the northeastern margin of the Berstein window, where the Middle Austroalpine Kirschlag klippen are in direct contact with the Penninics (Fig. 5.14). In the subsurface, to the E and NE of the Rechnitz metamorphic core complex the detachment fault can be traced downdip with confidence based on seismic data in Hungary described above. To the N and NW the upper/lower plate contact is placed on the northeastern flank of the Wechsel window (Fig. 5.13c). The isolated Tertiary basin fragment at Kirchberg (Ebner et al., 1991b) lies right on top of the Rechnitz detachment fault.
Farther to the NW and W the surface trace of the detachment fault is ill-defined; it dips probably gently to the N and runs between Lower Austroalpine units. The map of Flügel and Neubauer (1984) indeed shows repetition of these units in the area of Mürzschlag, which might be attributed to the detachment fault. To the W, the left-lateral Badenian Trofaiach Line (Nievoll, 1985) seems to crosscut the extensional system. While the upper/lower plate contact is clearcut along the Graz Paleozoic on the surface to the S, it is only tentatively placed at the northern margin of the Pohorje Mts., in the Remschnigg area (Fig. 5.13c).

In the subsurface the upper plate is identified as the Paleozoic succession of the Upper Austroalpine as described by Kröll et al., (1988). Unfortunately the contour of the upper plate beneath the Styrian Basin is uncertain. In my opinion the upper plate probably covers the central part of the basin as well in the Gnas subbasin, where there is no well control and Kröll et al. (1988) only inferred the subcrop of crystalline rocks (Fig. 5.13c).

The detachment fault itself was penetrated by a number of wells (Fig. 5.13c). In the Kainach Gosau area the Aßling U 1 well reached the Middle Austroalpine below the Graz Paleozoic (Kröll and Heller, 1978), whereas in the Styrian Basin wells Pichla 1, Wiersdorf 1 and Waltersdorf 1 reached the crystalline basement (Middle or Lower Austroalpine?) beneath Upper Austroalpine low-grade metamorphics (Kröll et al., 1988).

Returning to the cross-sectional geometry of the Rechnitz metamorphic core complex (Fig. 5.16), the detachment fault plane which is about subhorizontal in the Rechnitz area becomes gradually steeper to the ENE (up to 20-30°), at least in the observable 4 s two-way travel time interval of reflection seismic data (about 12 km depth). This steepening is in agreement with the model of Buck (1988), who proposed this geometry of core complex-forming detachment faults based on numerical simulations. Note that beyond an inflection point the detachment fault is postulated to flatten out in the middle crust. This geometry is similar to qualitative models proposed by Wernicke and
Axen (1988) and Lister and Davis (1989), but it is in sharp contrast to earlier models of Wernicke (1981, 1985) who considered this type of low-angle normal faulting to penetrate the whole lithosphere.

Note that the cross-section specifically implies that the Rechntiz detachment fault soles out in the middle crust (Fig. 5.16). Such a midcrustal level could coincide with the top of the European plate. The European basement promontory of the Bohemian Massif (Fig. 5.1) indeed reportedly projects below the eastern end of the Eastern Alps and can be followed to the SE below the Alpine edifice as far as the Graywacke Zone (Wessely, 1987, 1990). The onlap relationships on the flanks of the Bohemian Massif (Fig. 5.17) show that it was a positive topographic feature throughout the Mesozoic and Paleogene at the southern margin of the European plate. I suggest that the present-day triangular shape of the outcropping Bohemian Massif should be extrapolated to the SE and this basement promontory actually continues beneath the Alpine edifice all the way to the Rechner area (Fig. 5.18).

The strike-section illustrates the along-strike geometry of the Rechner detachment fault (Fig. 5.18). Note that although the detachment fault plane dips away from the Rechner core complex it still climbs up-section with respect to the footwall (or lower plate) both to the N and to the S. To the N the extremely low-angle Rechner fault cuts Eoalpine thrusts, which might be responsible for the repetition of Lower Austroalpine units in the Mürzzuschlag area (Fig. 5.13c). The fault is postulated to sole out at the base of the Austroalpine nappe-system. To the S the detachment fault may be identified as the low-angle normal of Plate 9, separating the Middle Austroalpine of the Pohorje Mts. from the Upper Austroalpine of the Zala Basin. Note that even in this strike section I infer the flow of the lower crust underneath the Rechner area (Fig. 5.18). The associated updoming of the lower crust should have occurred beneath the Rechner area (cf. Fig. 5.16) since the normal offset of the Rechner detachment fault has a local maximum there.
Fig. 5.17. Cross-sectional geometry of the Bohemian Massif in Austria from Wessely (1987).
CONCEPTUAL CRUSTAL STRIKE SECTION ACROSS
THE RÁBA RIVER EXTENSIONAL CORRIDOR

NORTHERN CALCAREOUS ALPS  SEMMERING  WECHSEL  RECHNITZ  EISENBERG
ZALA BASIN

UNEXTENDED UPPER PLATE
METAMORPHIC CORE COMPLEX
(LOWER PLATE)
DETACHMENT FAULTS OF WIDE-RIFT STAGE
(UPPER PLATE)

MOHO ~50 KM

UPPER AUSTROALPINE  LOWER AUSTROALPINE  FLYSCH AND HELVETIC
MIDDLE AUSTROALPINE  PENNINIC  EUROPEAN MESOZOIC

EUROPEAN CRYSSTALLINE  FLOWING LOWER CRUST

LEGEND

Fig. 5.18. Conceptual crustal strike section across the Rába River extensional corridor. Vertical exaggeration is about twofold.
and apparently diminishes to the N and S. That is why only the deepest tectonic units were unroofed in the central area.

I further propose that there should be a fortuitous relationship between the location of the Rechnitz metamorphic core complex and the underlying Bohemian basement spur. According to the recent work of Buck (1993), areas where low-angle normal faults occur are characterized by high heat flow and large crustal thickness. It is hard to evaluate the heat flow at the time of extension but it was likely to have been at least as high as the recently observed maximum values in the NW Pannonian Basin (90-95 mW/m²; Dövényi and Horváth, 1988). As to crustal thickness, at present the Mohorovicic discontinuity is at 35 km depth beneath the Bohemian Massif (e.g. Posgay et al., 1991) and this value is accepted as the paleothickness of the promontory throughout the Mesozoic and the Paleogene. During the Early Miocene the eastward escaping Alpine edifice (see below) overrode the promontory which resulted in a thickening of the crust by an additional 15-20 km, since the base of the escaping unit was most probably within the Penninic unit (cf. Ratschbacher et al., 1991). Such an anomalously thick (50-55 km) and possibly hot crust involving the Bohemian Massif provided the conditions (Buck, 1993) for the formation of a major NE-dipping normal fault. The prominent northeastern edge of the crystalline Bohemian Massif (Fig. 5.17) provided the stiff tip where the major normal fault could have nucleated. To the NE the Rechnitz fault was decoupled at midcrustal depth (15-20 km) at the inherited detachment surface on top of Europe. In the nearby Vienna Basin (Fig. 5.1) Neogene faults supposedly flatten out at the same surface, but at a lesser depth (6-8 km, see Royden et al., 1982). This view, however, is not accepted by Wessely (1988).

The original average dip of the Rechnitz fault can be estimated because from the lower plate Wechsel area the youngest ⁴⁰Ar/³⁹Ar ages were reported to be Cretaceous (Dallmeyer et al., 1992). Assuming that the breakaway for the Rechnitz fault was indeed in
the Koralpe region, some 90 km to the WSW, the original dip of the detachment fault was 6-7° (average!), following the logic of Richard et al. (1990), John and Foster (1993) and Dokka (1993). Along another profile across the Rechnitz area, the initial dip might have been larger since all the 40Ar/39Ar ages are reset (Kubovics, 1983; Koller, 1985). Note again, this type of calculation assumes an originally planar low-angle normal fault (cf. Buck, 1988; Wernicke and Axen, 1988).

Presently the Mohorovicic discontinuity is flat at about 30-32 km depth along the dip cross section (Fig. 5.16; Posgay et al., 1991), except at the western end where the Moho starts to deepen due to the Alpine root. The flat Moho means that space conservation requires the hot lower crustal material to flow laterally under the Rechnitz core complex. Such a mechanism is quite plausible as it was shown recently by Block and Royden (1990). Interestingly enough, this type of lower crustal "diapirism" is similar to structural features produced by salt tectonics (A.W. Bally, pers. comm.), obviously with important differences in terms of scale and velocity.

The above described core complex scenario is also supported by radiometric dating. A whole rock sample from the Hungarian part of the Berstein Mts. (Vashegy) provided 12 Ma K/Ar age (Kubovics, 1983). From the Austrian side of the Rechnitz window group Koller (1985) published K/Ar ages spanning the 22-19 Ma time period. Recently Dunkl (1990, 1992c), Demény and Dunkl (1991), Dunkl and Demény (in press), I. Dunkl (written comm.) reported fission track ages from not only the Rechnitz window group (Fig. 5.15a) but also from the Wechsel window (Fig. 5.13a). In the Rechnitz and Berstein windows these authors found zircon FT ages between 21.5 and 13.3 Ma. Within the Rechnitz window itself these ages display a definite younging to the E, which agrees with the core complex interpretation (i.e. gradual unroofing of deeper Penninic levels, see Fig. 5.15a). Apatite ages from the Rechnitz window group and the Wechsel window span the 6-15 Ma interval (Fig. 5.13a). This is very important, since as it was pointed out by Dunkl
(1992c): "the formation of these two structures is connected to the same extensional stage". This is not surprising since the Wechsel area is also located above the midcrustal projection of the Bohemian Massif (Fig. 5.18) and its synchronous uplift with the Rechnitz window group is in accordance with its lower plate position with respect to the Rechnitz detachment fault.

In addition, Dunkl (1992c) published FT ages from Middle Miocene syn-rift clastics. The stratigraphic age of the Sinnersdorf conglomerate outcropping in the Rechnitz area is Karpatian (17.5-16.5 Ma). From the components of this conglomerate Dunkl (1992c) found a very well-defined zircon age group with a 58-51 Ma peak, whereas the apatites formed a population with a mean of 13-14 Ma. The lack of Miocene zircon ages shows that until the Karpatian the detachment faulting did not unroof more than the upper 4-6 km thick layer of the lower plate (zircon blocking temperature: ~200°C). This is further confirmed by the fact that in these Karpatian clastics not a component of Penninic material was ever found (Pahr, 1980). Thus the dominant Lower Austroalpine material in the conglomerate records an earlier Paleocene/Eocene uplift event. As to the apatite ages which are younger than the stratigraphic age of the sediments by about 3 Ma, one should consider an age reduction due to post-depositional reheating. The cause of this reheating was most probably the widespread Badenian volcanism in the Styrian Basin (e.g. Sachsenhofer, 1991; Ebner and Sachsenhofer, 1991).

The classical Middle Miocene Styrian unconformity outcropping at several places to the E of the Sausal area (Fig. 5.13c) is characterized by a prominent angular discordance (Friebe, 1991a; 1993). This unconformity was dated by Friebe (1991b) as uppermost Karpatian (about 16.5 Ma, see Appendix B) and seems to postdate most of the core complex type extensional deformation in the Styrian Basin.

Currently available paleomagnetic measurements offer yet another independent data set, which unfortunately is not decisive for the core complex interpretation of the Rechnitz
window group. Vertical axis tectonic rotations inferred from paleomagnetic measurements by Márton et al. (1987) and Mauritsch et al. (1991) were ignored by Ratschbacher et al. (1990) and this led to discussion on the credibility of available paleomagnetic data (see the comment of Márton and Mauritsch, 1991 and the reply by Ratschbacher and Behrmann, 1991).

For future paleomagnetic studies in the Rechnitz area it is interesting to note that in the analogous Colorado River extensional corridor (see Chapter 8) Wells and Hillhouse (1989) documented large vertical axis rotations in extensional allochthons (up to 50°, both clockwise and counterclockwise). In contrast, Faulds et al. (1992) found significant horizontal axis rotations in highly tilted upper plate blocks in northern Arizona. These rotational mechanisms are clearly due to the large-scale detachment faulting, and by analogy they might be responsible for the declination and inclination anomalies reported by Márton et al. (1987) and Mauritsch et al. (1991) from the Rechnitz and Wechsel areas.

Another structural element which has been mapped seems to be also important in the broader Rechnitz area, i.e. the well-developed corrugations of the detachment surface parallel to the slip-direction with wavelengths of 5-10 km and amplitudes of 300-500 m (Fig. 15b). The amplitude of these corrugations was derived from a cross-section drawn perpendicular to the slip direction by Pahr (1980). These mullion or megagroove structures are quite characteristic for the lower plates of metamorphic core complexes (e.g. John, 1987).

The tectonic denudation and thinning of the upper plate in the Rechnitz-Sopron area can also be indirectly deduced from the sedimentary record (Tari, 1993), since changes in clast types of syn-rift sediments has been proved to be useful to infer the progressive unroofing of metamorphic core complexes (e.g. Miller and John, 1988).

The initiation of low-angle normal faulting in the Sopron area (Fig. 5.3) can be dated by outcropping breccias, conglomerates and sandstone formations which
accumulated above the fault planes. The earliest clastics are the Karpatian (17.5-16.5 Ma) Sinnersdorf Gravel of Austria (Pahr, 1984) and Ligeterdő (Auwald) Gravel of Hungary (Deák, 1981). At the base of the 600 m thick sequence the clast types are dominated by the rocks of the Upper Austroalpine (Ligeterdő Member). Higher in the section (Hochriegel Member) the crystalline rocks of the Lower Austroalpine Grobgneiss series are the most common clasts. In the uppermost part of the sequence (Brennberge Member) Wechsel crystalline rocks are also abundant, suggesting the exhumation and erosion of detachment faults within the highly thinned upper plate of the Rechnitz fault. The diameter of boulders in this member ranges up to 2 m, suggesting their deposition in mass-flow-dominated alluvial fans (Janoschek, 1931). The continuing fault activity is recorded by the boulders of the overlying 40 m thick Rust Gravel (Badenian; 16.5-13.8 Ma). The frequent appearance of boulder-size clasts in the whole sequence suggests formation of significant relief in a subaerial environment.

The amount of WSW-ENE extension across the Rechnitz detachment is difficult to constrain. The striking correspondence between the contours of the upper plate segments separated by the Wechsel and Rechnitz areas (Fig. 5.13c) suggests 40±5 km extension. Since this separation probably occurred during the advanced stage of detachment faulting when the lower plate updomed, this figure is clearly a minimum estimate. Additional extension within the upper plate can be estimated from the distance between the breakaway in the Koralpe and the western edge of the upper plate (30±10 km). The total of about 70 km extension is certainly too low, since it does not involve the pronounced early Middle Miocene thinning of the upper plate, especially in the Rechnitz area.

The best way to estimate the extension is to find piercing points within the lower and upper plate. The contact between the Lower and Middle Austroalpine can be traced fairly well within the lower plate in the St. Radegund area (Fig. 5.13c). The matching boundary can be found in the subsurface of the Danube Basin, described earlier (see
Panels 4-7). Their apparent offset is 80±10 km measured parallel with the transport direction of the upper plate. The problem with this estimate is whether the seismically mapped boundary indeed corresponds to its inferred tectonostratigraphic position? This uncertainty is the main reason for the error bar.

At any rate, in Chapter 6 a conservative estimate (70 km) will be used for a regional reconstruction. It should be emphasized, however, that the real magnitude of E-W extension across the Rába River extensional corridor could have been significantly more (for an indirect argumentation see Chapter 9).

5.4.1.3.3 SYN-RIFT MAGMATISM

The Middle Miocene volcanics of the Styrian Basin (Fig. 5.13b) are usually not mentioned in the discussion of the coeval volcanics of the Carpathian region (e.g. Szabó et al., 1992). Although the Styrian volcanites have a generally more alkaline character than their Carpathian counterpart (i.e. trachytes, shoshonites), Ebner and Sachenshofer (1991) speculated on their subduction-related origin. Based on the proposed shallow depth of the European foreland beneath the Rába River extensional corridor (10-20 km, see Fig. 5.18 and Plate 10) the melting of the "subducting" European plate at supposedly 100 km depth can be excluded as the source of magmatism.

Instead, the origin of the Styrian volcanics is better understood in the context of the above outlined core complex type extension. The rise and decompression of lower crustal material underneath this extensional terrain (Figs. 5.16 and 5.18) may induce melting of lower and middle crustal rocks. Since trace element studies are not available from the Styrian volcanics, this hypothesis remains to be tested.

5.4.1.4 EARLY MIOCENE ESCAPE TECTONICS

Regarding escape tectonics perhaps the most important consequence of this study is
that there appears to be no major left-lateral strike-slip fault (offset ~450 km) crossing the NW Pannonian Basin (e.g. Rába Line of Kázmér and Kovács, 1985). Throughgoing Middle Miocene sinistral faults can be found only at the edge of the Vienna Basin (Fodor et al., 1990; Fodor, 1991; cum lit.) and at the southern margin of the Northern Calcareous Alps (SEMP Line of Linzer et al., 1992). The postulated but widely accepted continuation of the Rába Line into the Hurbanovo-Diósjenő Line (e.g. Balla, 1988) should also be rejected.

5.4.2 MESOALPINE EVOLUTION OF THE STUDY AREA

During the Early Miocene the present-day Danube Basin was an elevated plateau (see Fig. 5.3). Sedimentological studies in Slovakia indicated a southern elevated area as the provenance of the thick Jablonica Formation clastic succession (Kovác, 1985, 1986; Kovác et al., 1989). A comparable formation (Csatka Gravel) is also known from the Bakony Mts. of Hungary, which is thought to be deposited in the fluvial environment of a high-relief terrain (Korpás, 1981). This high plateau seems to be the direct eastern continuation of the Augenstein Plateau of the Eastern Alps (Tollmann, 1986).

The only seismically resolved manifestation of this structural stage could be the enigmatic clastic wedge at the southeastern flank of the Danube Basin. This wedge consists of continental sediments (conglomerates, redbeds) that were regarded as Karpatian by Teleki et al. (1989) and Mattick et al. (1989). I accepted this age and therefore show them as part of the syn-rift sequence in Fig. 5.6. The peculiar geometry of this clastic wedge (i.e. thickening away from the basin) suggests that it may be older than Middle Miocene. If this is indeed the case then these Early Miocene sediments were deposited on the flank of the Csatka-Augenstein Plateau.

In the study area, the Paleogene succession was not studied because the Eocene is erosionally truncated just to the S of the area covered by the seismic grid (see Fig. 5.10).
Therefore the interpretation presented below is based on the reinterpretation of available stratigraphic, sedimentologic, radiometric (Chapter 4) and structural data involving a broader region than the NW Pannonian Basin (Fig. 5.19).

5.4.2.1 MIDDLE EOCENE - EARLY MIocene BASIN EVOLUTION

The onset of sedimentation in the Paleogene Basin of W-Hungary was diachronous (Fig. 5.20). Prior to the transgression from the SW (Dudich, 1977; Dudich and Kopek, 1980) at the beginning of the Lutetian (Fig. 5.21a), the area of the Bakony Mts. experienced a long period of subaerial exposure which is typical for much of the Alpine realm ("Paleocene restoration" of Trümpy, 1980). Bauxite deposits were formed on this continental terrain, which had a gentle NW-dipping paleosurface ( Mindszenty et al., 1988). The subcrop map at the basal unconformity ( Császá et al., 1978) shows the NE-SW striking trend of the underlying Mesozoic formations. The youngest formation below the basal unconformity is Maastrichtian in age ( Polány Marl) in the NW part of the Transdanubian Central Range. To the SE, the stratigraphic gap between the Eocene and the underlying Senonian is increasing and the oldest formation, the Csehbánya Formation, is Coniacian-Santonian in age. Along the edge of the Senonian basin, even older rocks are found immediately below the basal unconformity, but the distribution of these units also reflects the post-Aptian and pre-Senonian phase of emergence and subaerial erosion.

I interpret this stratigraphic gap which clearly increases towards the SE and the NE-trending Senonian formations below the basal unconformity as the result of erosion of a peripheral bulge. The karst bauxites formed on this uplifted bulge are poorly dated but are considered to be Late Paleocene-Early Eocene in age (Mindszenty et al., 1988). Their formation was controlled by minor NNE-SSW striking normal faults, which I interpret as structures due to "flexural extension" on the forebulge, commonly observed on the
Fig. 5.19. Index map of the Hungarian Paleogene Basin showing the locations of different geographic names mentioned in the text (after Balogh, 1971; Vass et al., 1979; Tanács and Rálisch, 1990). Location of Figs. 5.23 and 5.24 is also shown.
Fig. 5.20. Stratigraphy of the Hungarian Paleogene Basin from Tari et al. (1993). This compilation is based on sources listed in text.
Fig. 5.21. a,b,c. Simplified Eocene paleogeographic maps and structure sections of the Hungarian Paleogene Basins. The area shown in these maps is identical to the area shown in Fig. 5.19. No palinspastic reconstruction has been made in the construction of these maps. Modified from Tari et al. (1993). The cross-sections are meant to be merely illustrations rather than actual geologic sections.
Fig. 5.21. d,e,f. Simplified Oligocene to Early Miocene paleogeographic maps and structure sections of the Hungarian Paleogene Basins. The area shown in these maps is identical to the area shown in Fig. 5.19. No palinspastic reconstruction has been made in the construction of this map series. Modified from Tari et al. (1993). The cross-sections are meant to be merely illustrations rather than actual geologic sections.
foreland flank of foredeeps (e.g. Bradley and Kidd, 1991). These interpretations suggest the presence of a SE-vergent thrust-fold belt to the NW of the South-Bakony Mts. (in present-day coordinates).

The transgression at the beginning of the Lutetian (NP14 nannoplankton zone, Báldi-Beke, 1984) was probably a response to this thrust load (Fig. 5.21a). The continuing tectonic subsidence created accommodation space for neritic limestones on a carbonate ramp. With the gradual acceleration of thrust-induced subsidence, this shallow-water nummulitic limestone (Szóc Limestone) eventually drowned and deeper-water glauconitic marls (Padrag Marl) were deposited in a starved flexural basin (Fig. 5.21a). The turbiditic sandstone intercalations with abundant tuffaceous material of the upper Padrag Marl (Bernhardt et al., 1988; Báldi-Beke and Báldi, 1990) indicate the proximity of the basin axis during the Early Priabonian.

Unfortunately, there is no direct structural evidence for a Middle Eocene thrust-front to the NW of the South-Bakony. This area lies beneath the Danube Basin. Along strike, in Slovakia, there are some NW-dipping reflectors in the pre-Tertiary basement (Tomek and Thon, 1988; Fig. 5.22), which may be interpreted as SE-verging thrust planes that were active during the Paleogene. Tomek et al. (1987) themselves also interpreted these reflections as Alpine thrust planes reactivated as low-angle normal faults during the Middle Miocene (reflection events J, F and G).

For the Bartonian, the same starved deep-water flexural basin (Fig. 5.21b) shifted farther to the ENE in the Transdanubian Central Range (Fig. 5.21b; cf. Dudich and Kopek, 1980). In the northeastern part of the Transdanubian Central Range, the progressive deepening coupled with humid climatic conditions (Vörös, 1989) promoted the deposition of thick coal seams (Dorog Coal) along the shallow-water rim of the foredeep. Note that the Piszke Marl may be in part a lateral equivalent of the Padrag Marl.

During the Priabonian, thin-skinned thrusts propagated to the SE (Fig. 5.21c). The
Fig. 5.22. Deep reflection seismic line across the Slovakian Danube Basin from Tomek (1987). For location see Fig. 5.1.
synsedimentary growth of these NE-trending, SE-verging submarine anticlines was documented by Fodor et al. (1992b). One of them, the Buda Anticline (Fodor et al., 1992b), was also a facies boundary separating the bathyal Buda Marl from the neritic Szépvölgy Limestone. In my interpretation, the Buda Anticline was the leading edge of the thrust belt (Fig. 5.21c).

Later, during the Early Oligocene the deposition of the euxinic Tard Clay (Fig. 5.21d) indicates an overall regression that initiated the separation of the Tethys (Eoparateethys) from the Mediterranean (e.g. Báldi, 1986). The Paleogene thrust-fold belt became exposed (Fig. 5.21d), as suggested by frequent coarse-grained sandstone intercalations in the Tard Clay and the appearance of abundant reworked Cretaceous-Eocene nannoplankton floras at the base of the NP23 nannoplankton zone (Báldi-Beke, 1977). Also during this time, Eocene strata were eroded in the internal parts of the thrust wedge ("infra-Oligocene denudation" of Telegdi-Roth, 1927). The uplift and subaerial exposure of the thrust belt were probably related to the gradual waning of thrusting. The basinwide regression appears to be the result of the uplift of the Hungarian Paleogene Basin (Fig. 5.21d).

Dramatic changes took place in the Middle Oligocene (Late Kiscellian, about 31 Ma). At the beginning of the NP24 nannoplankton zone, the Hungarian Paleogene Basin was subsiding rapidly. Deposition of the euxinic Tard Clay ceased and normal marine connections were re-established in the bathyal sea of the Kiscell Clay (Fig. 5.21e, Báldi, 1986). At the basin margins, pronounced transgression occurred with the basal Hárshegy Sandstone, directed towards the NW. Fig. 5.21e shows the very particular spatial pattern of the transgression of the Late Kiscellian sea.

The unique geometry of this transgression is explained as follows. During the Middle-Late Eocene and the Early Oligocene, the thrust belt was characterized by NE-trending thinned-skinned thrusts (Fig. 5.21c,d), but at the end of the Early Oligocene, a
basement-involved thrust (Diósjenő-Hurbanovo(Ógyalla) Line, Fig. 5.21e) formed with a different, ENE-strike (Figs. 5.23 and 5.24). The Diósjenő Line is documented in the pre-Tertiary basement of the Pannonian Basin and is covered by Miocene sediments and volcanics. The Diósjenő Line is traced on the basis of borehole data (e.g. Dank and Fülöp, 1985; Fusán et al., 1987) and significant magnetic and gravity anomalies which are associated with it (e.g. Balla, 1988). The Diósjenő Line is a NW-dipping, steep structure of uncertain tectonic origin, according to Balla (1988). Since it separates the crystalline mass of the Veporides in the N from the non-metamorphosed Mesozoic of the Transdanubian Central Range in the S, it might well be a major thrust fault involving the basement. Thrusting with a new, deeper décollement level cross-cutting earlier thin-skinned thrusts is quite common in folded belts (see e.g. Wyoming; Bally, 1989).

If correct, then the particular outline of the Late Kiscellian transgression to the N of Budapest can be understood as the result of superposition (Fig. 5.21c) of the NE-trending fold-belt and its toe-trough (Fig. 5.21a) by a new ENE trending basement involved thrust and its foredeep trough (Fig. 5.21b).

The sands of numerous turbidite intercalations in the lower half of the Kiscell Clay derived from the foreland flank of the underfilled foredeep and were transported longitudinally to the NE-ENE along the axis of the foredeep. On the tectonically active side of the foredeep, coarse clastics derived from the emergent Vepor-unit are areally restricted (Báldi, 1986).

The pronounced transgression at the base of the Kiscell Clay suggests that tectonic processes may be the main reason for local transgressions in flexural basins. Transgressions such as that of the basal Hárshegy Sandstone do not necessarily indicate eustatic sea-level rises (Lakatos et al., 1992) and, in this particular case, tectonics apparently overprinted any third-order eustatic signal (Tari and Sztanó, 1992).

During the Late Oligocene, widespread fluvial deposition began in the subaerially
Fig. 5.23. Map view of the Hurbanovo-Diósjenő Line from Tari et al. (1993). The outline of this map is indicated in Fig. 5.19.
Fig. 5.24. Cartoon depicting the bathymetric changes in the central part of the HPB from Tari et al. (1993). The outline of this map is indicated in Fig. 5.19. a) Idealized map view of the "Buda Line" with the corresponding toe-trough during the Late Kiscellian, before the beginning of thrusting along the Diósjenő Line. b) Idealized map view of the Diósjenő Line right after the beginning of thrusting during the latest Kiscellian (~31 Ma), with the supposed extent of the corresponding toe-trough. Note that I assumed a relatively shallow and narrow foredeep for a) and a deeper and wider foredeep for b) based on the thin-skinned vs. thick-skinned character of the Buda "Line" and Diósjenő Line, respectively. c) The superposition of the two foredeeps gives the same pattern as shown in Fig. 5.21e.
exposed Transdanubian Central Range (Csatka Gravel). Sedimentological studies indicate a highland in this area (Korpás, 1981), which was the eastern prolongation of the Augenstein plateau of the Oligocene Eastern Alps (e.g. Tollmann, 1986). I interpret this elevated plateau as the result of isostatic uplift after the cessation of thrusting in this part of the Paleogene Basin. To the E, coarse clastics were deposited at the basin margins (Törökbálint and Eger Sandstones; Fig. 5.17). In the axis of the basin deposition of the Szécsény Siltstone indicates shallower water depth (200-400 m, Báldi, 1986) than the underlying Kiscell Clay.

The Late Oligocene-Early Miocene is still characterized by very high subsidence rates (Báldi, 1986) and also asymmetric thickness distribution of the different lithostratigraphic units. Compared to the Middle Eocene-Early Oligocene, the markedly increased deposition rates during the Late Oligocene - Early Miocene (Báldi and Báldi-Beke, 1985) are interpreted as the expression of the overall change from the "flysch" to the "molasse" stage (e.g. Allen et al., 1991) or in other words from the "underfilled" to the "overfilled" stage in the evolution of the flexural basin (Covey, 1986). The regressive character of the whole Late Oligocene-Early Miocene succession of the Hungarian Paleogene Basin (Fig. 5.17) also indicates its overfilling.

The Lower Miocene (Eggenburgian) neritic formations of the Hungarian Paleogene Basin (Pétervására and Budafok Sandstones; Fig. 5.21f) show a definite basinward shift of facies (Sztánó and Tari, 1993). Despite this regressive trend, the deposition of the bathyal Szécsény Siltstone continued in the basin axis (Fig. 5.21f) until the beginning of the NN3 nannoplankton zone (~19 Ma).

Finally, at the end of the Early Miocene, isostatic uplift following the cessation of thrusting combined with the increased amount of sediment supply (Budafok Sandstone) resulted in the cessation of marine sedimentation. Arealy extensive continental clastics (Zagyvapálfalva Clay) covered the uplifted and eroded strata of the former foredeep. All
these features are characteristic for flexural basins (Heller et al., 1988).

The Lower Miocene (Ottnangian) Gyalakeszi Rhyolite Tuff fell on an already subaerially exposed region (Hámor, 1985), providing a regional isochron, which we consider as the beginning of an entirely different, new geodynamic episode. The volcanic activity which produced the Gyalakeszi Rhyolite Tuff at the beginning of the Ottnangian (Hámor, 1985; Fig. 5.20) and the extensive Lower Miocene ignimbrite units (K/Ar ages: 19-17 Ma, Hámor et al., 1987) exposed to the S of the Bükk Mts. (Fig. 5.19) are in my interpretation related to the right-lateral transpressive movements along the Balaton-Mid-Hungarian-Periadriatic shear zone. I consider this time interval as the beginning of large-scale continental escape (cf. Kázmér and Kovács, 1985) of the North Pannonian unit from the Alpine realm, which resulted eventually in the back-arc extension of the Neogene Pannonian Basin (Tari et al., 1992b).

5.4.2.2 PALEogene MAGMATISM

Paleogene volcanic centers are known in several areas of the Hungarian Paleogene Basin (Fig. 5.19). In the Zala Basin volcanics (Szentmihdly Andesite) were found in wells, in the area of Bak and Zalaszentmihály (Körössy, 1989). These are Priabonian in age and coeval with another volcanic center in the Velence Mts. (Nadap Andesite). In the latter area, andesite dykes of Middle Oligocene age were also found (Darida-Tichy, 1987).

Petrological studies on upper Eocene-lower Oligocene magmatic rocks have shown that these predominantly calc-alkaline suites are indeed associated with a subduction process (Darida-Tichy, 1987). More specifically, their geochemistry suggests an intermediate position between an active continental margin type and an oceanic island arc type volcanism (Darida-Tichy, 1987). The fact that this magmatism occurred within the "retroarc" basin may indicate that the angle of Paleogene subduction was relatively shallow (see Chapter 6 for a regional geodynamic model).
5.4.3 EOALPINE EVOLUTION OF THE STUDY AREA

In this subchapter, I describe first the distribution, internal seismic facies, structure and tectonic interpretation of the Senonian sequence. In order to discuss the compressional structure of the pre-Senonian basement it is unavoidable to summarize the potential décollement levels in the Paleo-Mesozoic Upper Austroalpine succession of the Bakony Mts. A set of dextral strike-slip faults appears to be decoupled on these surfaces.

Several thrust sheets will be described for the southeastern part of the Hungarian Danube Basin and the Zala Basin. These features now seem to be connected with the well-known thrusts of the Balaton Highland forming a newly defined Eoalpine thrust-fold belt. The alternative explanations offered to resolve the large-scale bending of this folded belt will be briefly discussed. Finally a concise summary on the Eoalpine evolution of the NW Pannonian Basin will be presented.

5.4.3.1 SEISMIC FACIES AND TECTONICS OF THE SENONIAN BASIN

Mapping the internal seismic facies of the Senonian basin in the Danube and Zala Basins and in the northwestern flank of the Bakony Mts. (see Fig. 5.2), a very strong correlation was found between the paleomorphology of the pre-Senonian basement and the distribution of reef-like features (see Panels 10 and 11). These buildups tend to be associated with the local highs (cuestas) of the basement (Fig. 5.25) and I interpret them as the reefs of the Campanian Ugod Limestone (Fig. 4.9). This supports the observations of Haas et al. (1986) that the lithofacies of the Senonian succession shows a good correlation with the underlying paleotopography (see Fig. 4.15). Here I propose that in fact the distribution of Senonian lithofacies is controlled by the geometry of the underlying Eoalpine thrusts (Fig. 5.25). Synsedimentary tectonics as normal block-faulting (e.g. Bignot et al., 1984) does not seem to be tenable to explain this relationship.

Additionally, I propose a specific basin-forming mechanism for the origin of the
Fig. 5.25. Effect of inherited paleotopography on Santonian (lower) and Campanian (upper) lithofacies. See Fig. 4.9 for Senonian lithostratigraphy.
Senonian Basin. The only documented evidence for syn-sedimentary tectonics is the intra-Senonian unconformity observed on several seismic sections (e.g. section D13). This unconformity appears to be local in character and it was identified by onlapping reflectors within the Polány Marl (see section D9 calibrated with the Dabrony-1 well).

Another hint of syn-sedimentary tectonics was reported by Haas (1979) who found thick (up to 100 m) breccia bodies embedded in the pelagic Polány Marl in the Bakony Mts. This Jákóhegy Breccia Member consists of neritic carbonate fragments derived from the nearby carbonate buildups of the Ugod Limestone. Although I could not demonstrate the direct link between the subsurface intra-Senonian unconformity and the outcropping Jákóhegy Breccia I propose that they were formed by the same tectonic mechanism.

Fig. 5.26 shows the most probable scenario for the formation of the Lower Maastrichtian Jákóhegy Breccia bodies (Góczán, 1964) within the Polány Marl and the intra-Senonian unconformity introduced in this thesis. Abandoned Eoalpine (i.e. Early to Middle Cretaceous) thrust planes were probably reactivated during a prominent compressional event which resulted in the downslope shed of neritic carbonate material. After the short-lived uplift of the frontal part of the thrust sheets, continuing deposition of the Polány Marl formed the observed onlaps on the flanks of drowned carbonate reefs. Interestingly enough in the coeval Lombard Basin (Fig. 2.18) a similar lithofacies (Missaglia Megabed) was documented in a pelagic sequence (Bernoulli et al., 1981). This latter lithofacies, however, is slightly older, Lower Campanian in age.

This interpretation challenges the traditionally held "post-tectonic" character of the Senonian Basin. Even if the Senonian succession apparently seals the Eoalpine overthrust contacts below, it cannot be considered as an inactive "epicontinental" basin (cf. Haas, 1979). It is proposed here that this basin can be best understood in terms of a flexural basin. The thick coal-bearing strata (Haas et al., 1986, 1992) characterized by Appalachian-type cyclotherms at the base of the succession, the appearance and sudden
Fig. 5.26. Speculation on the origin of the intra-Senonian (Maastrichtian?) unconformity. Note that Senonian formations underlying the Polány Marl are not shown, cf. Fig. 5.27.
drowning of a carbonate platform and the subsequent deepening to bathyal water depth is an asymmetric facies cycle, quite typical for the foreland flank of flexural basins (Tari, 1992b).

An indirect argument for the flexural origin of the Senonian Basin is the prominent analogy between the lithofacies development of the Eocene, Senonian and Middle Cretaceous basins of the Transdanubian Central Range (Fig. 5.27). Note that the column of the Early-Middle Cretaceous basin is an idealized column, since many of the formations shown are stacked together laterally rather than vertically in reality, representing mainly heterotopic facies assemblages (see Fig. 4.9).

At any rate all these basins are underlain by important bauxite deposits (Fig. 5.27) as pointed out by Mindszenty (1988). Transgressive basal clastics follow in all cases. Major coal seams were deposited in the Senonian and Eocene Basins, a feature thought to be characteristic for the foreland flank of foredeeps (McCabe and Parrish, 1992). The coal, however, is not significant at the base of the Early-Middle Cretaceous basin. Continuing deepening resulted in carbonate deposition in all these basins. The carbonates, however, formed a wide carbonate platform in the Eocene Basin, whereas carbonate buildups in the Senonian and Early-Middle Cretaceous basins were probably more influenced by the inherited paleotopography. All these neritic carbonate successions have a sharp boundary with overlying pelagic marls. The condensed section observed in some cases on top of this surface has a resemblance to the basal foredeep unconformity of Bally (1989). The rather monotonous pelagic marl sequence gradually passes into either basin plain turbidites (Eocene and Senonian Basins) or into a flysch succession (Early-Middle Cretaceous basin). In the latter case even the uppermost part of the flexural basin was preserved recording the approach of the thrust-fold belt in forms of proximal fans (Sztánó, 1990).

A unique feature associated with the Senonian Basin is the appearance of lamprophyre dikes (Horváth and Ódor, 1984) and red calcite veins (Haas et al., 1985) just
Fig. 5.27. Cretaceous-Eocene flexural basins of the study area.
before the transgression of the Senonian sea (Coniacian/Santonian). These dikes were previously regarded as the manifestation of extension (Kázmér and Szabó, 1989). I more specifically propose that they were formed as the result of flexural extension on the forebulge of the approaching Senonian Basin. Extension on the forebulge of flexural basins is quite common (see Hancock and Baven, 1987; cum. lit.).

To this end it is important to emphasize that the Senonian Kainach Basin some 100 km to the W (5.13a) was interpreted recently as related to unroofing extension (Fig. 5.2) at high crustal levels in the Alpine orogenic wedge (Krohe, 1987; Ratschbacher et al., 1989; Fritz et al., 1991). This interpretation seems to be in apparent conflict with the flexural model outlined above for the Senonian Basin of Transdanubian Central Range. I will address this problem in Chapter 7.

5.4.3.2 DÉCOLLEMENT LEVELS IN THE PALEOZOIC AND MESOZOIC OF THE BAKONY MTS., DANUBE AND ZALA BASINS.

The Middle Cretaceous compression deformed a thick succession of Paleozoic-Triassic-Jurassic-Neocomian rocks. Since this stratigraphic column is made up of very heterogeneous lithologies, certain levels are more likely to have acted as décollement levels while other, more competent levels provided ramps of the compressional structures.

Several documented cases of repeated sections in outcrops and boreholes interpreted as thrusting have been published in the area of the Bakony Mts., and the Danube and Zala Basins. The compilation of these examples in Appendix A.6 provides a first attempt to identify potential Eoalpine décollement levels in the Paleozoic-Mesozoic column.

The stratigraphic position of major Mesozoic décollement levels is shown in Fig. 5.6. Note that these levels are comparable to those described from the Southern Alps (e.g. Laubscher, 1985; Roeder, 1989; Doglioni, 1992; Schönborn 1992) or from the Northern Calcareous Alps (e.g. Laubscher, 1989; Eibacher et al., 1991; Linzer et al., in press).
5.4.3.3 EoALPINE STRIKE-SLIP (TEAR?) FAULTS

The Eoalpine structures are crosscut by a number of WNW-trending, steeply dipping faults in the southeastern part of the Hungarian Danube Basin (Fig. 5.4d). Following them to the surface exposures of the Bakony Mts., they became right-lateral strike-slip faults (see Fig. 5.2). Some of these faults are documented on several consecutive seismic lines. Panels 10 and 11 show the subsurface extension of the Telegdi-Roth and Padrag Lines, respectively. Panel 14 shows a new strike-slip fault which I termed Nagytílaj Fault (Fig. 5.4d), after the Nagytílaj(Nt)-2 borehole, allowing for dating the strike-slip movements (see subchapter 5.2.10). The Cretaceous Nagytílaj Fault postdates the Barremian Sümeg Marl and predates the basal Senonian (Santonian) strata. By inference I place the strike-slip activity in the early Albian (see next subchapter).

All of these strike-slip faults of the Transdanubian Central Range (e.g. Telegdi-Roth, 1935; Mészáros, 1983; Balla and Dudko, 1989; Gyalog, 1992) were compiled in a map shown in Fig. 5.28. Apparently the faults are restricted for the upper level of the Upper Austroalpine system, suggesting their decoupling at depth within this nappe system. This is the logical explanation (Tari, 1991) for the well-documented termination of some of the strike-slip faults at the Veszpém or Litér thrusts in the Balaton Highland.

These Middle Cretaceous strike-slip faults (reactivated during the Miocene) display a close structural resemblance to those recently described for the Northern Calcareous Alps of Austria (Eisbacher et al., 1991; Linzer et al., 1992; Linzer et al., in press; Eisbacher and Brandner, in press) as it was pointed out by Tari and Horváth (1992). Quite importantly this suggests a once continuous Upper Austroalpine cover connecting the Northern Calcareous Alps with the Bakony Mts. (see Chapter 6). Eisbacher et al. (1991) proposed that the wrench faults in the Northern Calcareous Alps are tear faults coeval with the Eoalpine compression in the Alpine folded belt. The same may hold for the faults of the Transdanubian Central Range (see next subchapter).
Fig. 5.28. Tectonic map of the pre-Senonian basement in the study area (cf. Fig. 4.3b).
5.4.3.4 EOALPINE THRUSTS AND NAPPEs

In this thesis I mapped the subcrop of the pre-Senonian basement in the Hungarian Danube Basin and the northern part of the Zala Basin (Fig. 5.4d). One of the most striking results of this mapping was that two NW-vergent thrusts seem to be identical to the SE-vergent Veszprém and Lítér thrusts in the Balaton Highland described in Chapter 4 (Figs. 5.29 and 5.30). Moreover, in the case of the Veszprém thrust it could be directly correlated with its counterpart in the Danube Basin following the base of the thick Hauptdolomit mass (Fig. 5.4d). I named the thrusts sheets in Fig. 5.29 after the wells which drilled into these structures (Fig. 5.30). The individual Eoalpine thrusts are documented in a number of seismic sections (Panels 4-15) in the Danube Basin.

Note that in addition to the NW-trending dip section the perpendicular strike section reveals SW-verging structures (Fig. 5.30). The geometry of Eoalpine structures in the Zala Basin (Fig. 5.29) should be regarded speculative because of the lack of adequate seismic data at present. Thus interpretation is based only on the general well descriptions of Körössy (1988). If correct, the main décollement surface for the SW-vergent structures is located at the boundary of the Veszprém Marl and the Hauptdolomit (Fig. 5.7).

The NW-vergent Eoalpine structures involving different levels of the Mesozoic section utilized other décollement surfaces as well, such as the base of the Middle Triassic and the base of the Lower Triassic (Fig. 5.7). Similar décollement levels were reported from the structurally analogous Krappfeld area (see Wolter et al., 1982; Appold and Pesch, 1984; Ratschbacher and Neubauer, 1989), located some 200 km to the W along strike (Fig. 5.1).

The deeper, mostly very low- to low-grade Paleozoic succession of the Upper Austroalpine is also characterized by NW-vergent thrusts (Fig. 5.4d). Along-strike correlation of paleotopography and analysis of distinct lithologies reported from wells in the Ikervár, Mihályi, Ölbö and Bükk areas led to a revised tectonostratigraphic subdivision
Fig. 5.29. Cartoon emphasizing the idealized map view character of Eoalpine thrusts in the study area. Compare Figs. 4.13, 4.19, 5.4d and 5.28. Traces of cartoon cross-sections shown on Fig. 5.30 are also indicated.
Fig. 5.30. Cartoon emphasizing the idealized cross-sectional character of Eoalpine thrusts in the study area. Compare Plates 4, 5 and 10. Location of cross-sections are shown on Fig. 5.29.
of the low-grade Paleozoic basement in Hungary (Fig. 5.31, cf. Fig. 4.5). Possible correlative Paleozoic lithologies in the Styrian Basin are also shown and they are described in Appendix A. An important corollary of the new tectonic map shown in Fig. 5.4d and the regional transect (Plate 9) is that Paleozoic lithological correlations should also be attempted with the structurally analogous Graywacke Zone (see Appendix A).

5.4.3.5 BENDING OF THE TRANSDANUBIAN CENTRAL RANGE

The northeastern part of the Transdanubian Central Range in the new tectonic map (Fig. 5.28) is based on the work of Dudko (1992a, see Fig. 4.13) since to date there are no seismic reflection data in that area. Dudko (1992a) used well information to suggest the gradual bending of Eoalpine structures, including the Litér thrust and the first-order syncline of the Transdanubian Central Range. This change in strike has been postulated for many years (see subchapter 4.8.1) and even if it is correct the formation of the bending is not yet understood. In the following I propose alternatives to the gradual bending model.

Fig. 5.32a shows the currently held opinion (e.g. Balla, 1988) in a simplified cartoon. Since this model is based on the conceptual interpretation of scattered well data (Balla and Dudko, 1989) an obvious alternative is the superposition of two thrust fronts (Fig. 5.32b). The NE-trending, SE-verging thrusts are well established in the Balaton Highland. Evidence for NW-trending, SW-verging thrusts was found in the course of surface mapping (Wein, 1977) in the Buda Mts.

Yet another possibility to explain the bended thrust contacts is that they manifest the axial projection of the NE-plunging Velence "pericline" (Fig. 5.32c, cf. Fig. 4.13). Along strike to the SW the regional transect (Plate 10) suggests a major anticline underlain by blind thrusts involving crystalline (Middle Austroalpine?) basement. The formation of the Velence anticlinorium is responsible for the first-order syncline of the Bakony Mts., and this major axial culmination projects below the Buda Mts. to the NE.
Fig. 5.31. Revised Paleozoic lithostratigraphy of Eoalpine thrust units in the central Danube Basin (cf. Fig. 4.5). See also Appendix A for regional correlation possibilities.
Alternative models for the "bending" of the Transdanubian Central Range

(a) Gradual Bending

(b) Superposition

(c) Axial Culmination

Fig. 5.32. Alternative models for the bending of the Transdanubian Central Range. Area covered by these cartoons is shown in Fig. 5.28. See text for detailed explanation.
Although I favor the latest scenario (Fig. 5.32c), the problem of "bending" in the northeastern part of the Transdanubian Central Range remains open due to lack of adequate subsurface data.

I summarized the conclusions on the Eoalpine evolution of the NW Pannonian Basin in Fig. 5.33. Note that the structures on these cartoons are shown in their restored pre-extensional position (see Chapter 6 and cf. Figs. 5.1 and 6.2).

The first Eoalpine compressional deformation is manifested by the rapidly deepening Neocomian (Berriasian-Barremian) trough in the Gerecse Mts. (Fig. 5.33a). The drowning of the Tithonian carbonate platform representing a passive margin sequence is due to the approach of the thrust front from the E closing the Vardar oceanic trough. The Neocomian sequence is traditionally regarded as eroded in the middle part of the Transdanubian Central Range (e.g. Haas and Császár, 1984; Fig. 4.9). I propose that this sequence is primarily missing due to non-deposition on a WNW-trending forebulge (Fig. 5.33a).

By the Aptian the tip of the Vardar foredeep propagated farther to the W (Fig. 5.33b) suggested by the onset of flysch deposition in the Gerecse Mts. (Lábatlan Sandstone, Fig. 4.9). At the same time contractual deformation began to propagate into the cover sequence of the foreland along décollement levels in the Upper Austroalpine. The SW-verging imbricates in the Zala Basin (Fig. 5.30) are attributed to this stage. This interpretation is supported by the microtectonic observations of many in the Upper Austroalpine Graz and Gurktal Paleozoic (Fig. 5.1) that the very first Eoalpine thrust movements occurred to the WSW (e.g. Ratschbacher, 1986; 1987; Neubauer, 1987; Fritz, 1988). This top-to-the-W movement was dated by radiometric methods and the results vary between 120-130 Ma (Kralik et al., 1987; Fritz, 1991). Note that from the Mihályi thrust unit (Fig. 5.30) 116-123 Ma K/Ar dates were obtained by Árkai and Balogh (1989), which in my opinion reflects the age of this first compressional stage. Older ages reported
Fig. 5.33. Summary of Eoalpine evolution of the NW Pannonian Basin. Note that the structures on these cartoons are shown in their restored pre-extensional position (see Chapter 6 and cf. Figs. 5.1 and 6.2). See text for a detailed explanation.
from other Eoalpine thrust units involving Paleozoic rocks (Fig. 5.31) are reset ages between the Hercynian regional metamorphism (~320 Ma) and the first Eoalpine compression (~120 Ma). As to the sedimentary record the widespread Aptian Tata Limestone (Fig. 4.9) appears to postdate this deformation.

By the Albian the earlier WSW-ENE directed compression was replaced by a NNW-SSE compression (Fig. 5.33c). This change apparently occurred gradually based on the smoothly changing translation paths of many Eoalpine nappes observed in the Eastern Alps (e.g. Ratschbacher, 1986; 1987). During the Albian the first-order syncline of the Transdanubian Central Range formed with the associated Velence anticlinorium. The probably inherited large-scale detachment at the base of the Hauptdolomit resulted in the multiple repetition of this rigid unit along the axis of the first-order syncline. The smaller scale geometry actually shows two second-order detached synclines floating on top of the rigidly deforming Hauptdolomit mass (Plate 10). The NW-SE shortening was partly accommodated by a set of dextral strike-slip faults which might have acted as tear faults.

The effect of the still active Vardar trough is evidenced by the proximal fans of the Köszörüköbánya Conglomerate in the Gerecse Mts. (Fig. 4.9). This foredeep was generally migrating to the present-day SW and was superimposed on the above outlined compressional style. During the Early Albian I assume the migration of the corresponding forebulge farther to the SW to the Bakony Mts. where the Alsópere Bauxite was deposited on it. The areal distribution of the overlying, mostly fluvial to neritic Albian sediments (Fig. 4.9) is restricted to the axes of the second-order anticlines (e.g. Császár, 1986) apparently postdating them. By the Cenomanian compressional deformation slowed down or temporarily ceased by the closure of the Vardar foredeep, and the whole area experienced uplift and subaerial exposure.

During the Santonian another flexural basin began to propagate into the NW Pannonian Basin from the SW (Fig. 5.33d). Based on the stacking pattern of successive
sedimentary facies the Senonian Basin of the Bakony Mts. represents the foreland flank of the flexural basin at its southeastern margin. The compression-driven deepening of this basin resulted in its gradual migration to the E. This migration, however, stopped near the Cretaceous/Paleocene boundary and the area was again uplifted and subaerially exposed.

Perhaps the most important new aspect of this subchapter is that the above described Eoalpine structures in the subsurface of the Danube Basin and in the Bakony Mts. display comparable Alpine structural style to their counterparts in the Eastern Alps. The sharp tectonic distinction between the Eastern Alps and the North Pannonian unit as it was introduced in Chapter 2 is clearly artificial. The basement of the NW Pannonian Basin is indisputably made up of Eoalpine nappes and thrusts, and the almost century-old opinion on the autochthonous nature of the Transdanubian Central Range (see Chapter 1 for details) should be abandoned. The results of this thesis fully confirmed the original ideas of Uhlig (1903), Pávay-Vajna (1931) and Rozlozsnik (1937) on the *allochthonous* character of the Hungarian Mid-mountains.

5.4.4 EARLY ALPINE EVOLUTION

The currently available data do not permit the specific reconstruction of the Jurassic or Triassic extensional history of the NW Pannonian Basin. Only a few attempts exist in the literature which attempted to outline the general characteristics of the Early Alpine evolution of the study area (e.g. Galácz et al., 1984; Galácz, 1988; Budai and Vörös, 1992, 1993).
CHAPTER 6

IMPLICATIONS FOR THE ALPINE EVOLUTION OF THE PANNONIAN BASIN AND THE SURROUNDING MOUNTAIN BELTS

This chapter presents a series of palinspastic sketch maps of the Alpine belts of Central-Eastern Europe and their subsurface continuation underneath the Pannonian Basin. Similarly to Chapter 5, the discussion here begins with the present-day situation and goes backwards in time. The reconstruction is attempted in several steps by moving large-scale tectonic blocks. Not only is the map-view area balance respected in the reconstructions but also the sometimes severe internal deformation of these blocks revealed by their cross-sectional geometry. The results of Chapters 3 and 5 are incorporated into this analysis to constrain better the still considerable margin of error in the reconstruction.

6.1 RETRODEFORMATION IN TERMS OF KINEMATICS

Several pre-Neogene kinematic markers were selected for the reconstruction (Figs. 6.1 and 6.2). These include a number of dismembered Paleogene basins with their pre-Neogene subcrops (Tari et al., 1993) and the outline of the Mesozoic Vardar ophiolites (Csontos et al., 1992). The present-day outline of the two major tectonic units in the intra-Carpathian area is also shown.

These kinematic markers and the present-day outcrop outline of several mountain ranges are also shown on the next two maps (Figs. 6.3 and 6.4) to give a legible display. In the following larger-scale maps (Figs. 6.5, 6.6, 6.7 and 6.8), however, the same kinematic markers are only shown in a simplified manner. This type of reconstruction stops at the Eocene, although earlier Eoalpine tectonic scenarios will be briefly discussed.
6.1.1 NEOALPINE (MIOCENE-RECENT)

During Neoalpine times two fundamentally important tectonic processes took place: the Middle Miocene syn-rift extension in the Pannonian Basin and immediately before that the Early Miocene continental escape of the Northern Pannonian unit. The Quaternary inversion of the basin is still in its initial phase and the associated apparently negligible compressional shortening is ignored in the following.

6.1.2 RECONSTRUCTION OF THE MIDDLE MIOCENE EXTENSION

The restoration of the pre-extensional geometry of the intra-Carpathian area became of primary importance for further reconstructional steps. A possible approach was taken by Csontos et al. (1992), who used an unpublished reconstruction of the West Carpathian flysch belt by Oszczytko and Slaczka. In this work I have estimated the extension within the back-arc basin rather than the compression in the thrust belt.

In the Hungarian part of the North Pannonian unit, the Danube Basin is characterized by low-angle detachment faults and the Rechnitz metamorphic core complex. The amount of extension in this area is estimated as 40 and 70 km, in a NW-SE and a SW-NE direction, respectively (Fig. 6.2, see results of Chapters 3 and 5). More moderate extension was observed to the E of the Danube, where 30 km seems to be a reasonable value for the extension magnitude in a W-E direction. In the eastern end of the North Pannonian unit extension seems to be large again. In the basement of the East Slovak (or Transcarpathian) Basin Penninic rocks were found recently (Soták, 1993), suggesting an extensional style comparable to that of the Danube Basin. Therefore the E-W extension was estimated as large as 100 km.

In the South Pannonian or Tisia unit, the southern part of the Great Hungarian Plain displays a highly extended terrain (see Chapter 3); therefore I assumed a large amount of extension (180 km) in a ENE-WSW direction, while the extension is less dramatic (40 km)
Fig. 6.1. Present-day geometry of kinematic markers in the Pannonian/Carpathian region.
Fig. 6.2. Pre-extensional geometry of kinematic markers in the Pannonian Basin.
Fig. 6.3. Pre-escape geometry of kinematic markers in the Pannonian Basin.
in the perpendicular direction. The 180 km of extension indicates a slightly higher crustal thinning factor (β=2.4) than was calculated by Royden (1988) and Horváth et al. (1988) (i.e. β=2.2). This drastic estimate is independently justified by the shape of the Vardar ophiolites displaying an apparently offset branch beneath the Transylvanian Basin. The necking of this branch can be removed only by restoring a large amount of extension within Tisia. To the SW of the Mecešek Mts., in the Drava Basin the amount of extension by low-angle normal faulting is estimated as about 50 km. The same value was assumed to be persistent along strike (Fig. 6.2), for the southernmost, Serbian part of the Pannonian Basin, in the region of Fruska Gora (for geography see Fig. 2.16b). Taking into account the Miocene K/Ar ages reported by Pamic (1989) from several metamorphic units in that area (e.g. Motajica Mts., see Fig 2.16b), a "core complex"-type extension indeed seems to be a reasonable assumption.

In addition, the Eastern Alps and the Dinarides as a whole were moved back to the SW by 40 km to compensate for the push exerted by the northward moving African promontory (Adria).

In the pre-extensional restored map (Fig. 6.2) Tisia is shortened to two-third of its original length. The North Pannonian block shows less severe deformation, although it is also shortened both along strike and dip. The northeastern part of the North Pannonian block was also rotated by about 15° clockwise around a rotation pole just to the S of the tip of the Bohemian Massif. This rotation is necessary to keep together the major blocks along the Mid-Hungarian Line.

It is important to realize that this type of reconstruction did not produce overlaps (cf. Balla, 1984) and the northeastern perimeter of different units forms a relatively smooth curve displaying the pre-extensional outline of the inner boundary of the Outer Carpathian thrust-fold belt (cf. Oszczypko and Slaczka, in Csontos et al., 1992).
6.1.3 RECONSTRUCTION OF THE EARLY MIOCENE ESCAPE

In the second reconstructive step (Fig. 6.3) the eastward movement of the North Pannonian unit was restored first. The amount of right-lateral displacement along the Periadiatric-Mid-Hungarian Line due to this escape movement was estimated by the offset Hungarian Paleogene Basin. This value is 310 km on the pre-extensional map (Fig. 6.2) between the eastern ends of the Hungarian and Slovenian Paleogene Basins. The whole North Pannonian unit was moved to the W by this amount. (It is to be noted that this large-scale strike-slip movement along the Periadiatric Line is not accepted by many Alpine geologists; for a detailed discussion see Chapter 8.) The trace of the previously bent Mid-Hungarian Line is straightened out and the North Pannonian Unit was further rotated clockwise by 15° around a rotation point close to Budapest (restored position).

At this point I make another important departure from previous reconstruction attempts (e.g. Balla, 1984; Csontos and Vörös, in press). All these workers rotate Tisia back in a counterclockwise manner, by 100° based only on paleomagnetic data from the Apuseni Mts. (Patrasu et al., 1990). The drawbacks of such a reconstruction are discussed in Chapter 8. Here I rotated Tisia clockwise by 40° and moved it to the S by about 100 km. Beside the advantages of this approach in terms of Mesozoic facies relationships (see later), in terms of kinematic markers the Szolnok Basin ends up in the continuation of the Central Podhale Basin, the triangular shape of the Vardar ophiolites is further simplified into a more stripe-like zone and it finds its continuation to the N beyond the future trace of the Mid-Hungarian Line in the ophiolites of the Bükk Mts (Fig. 6.3).

In this reconstruction again the northeastern perimeter of different units forms a relatively smooth curve displaying the pre-escape outline of the inner boundary of the Outer Carpathian thrust-fold belt.

This type of reconstruction cannot be continued backwards without considering a much wider region, i.e. the whole Central Mediterranean. Moreover, not even the
approximate amount of Cretaceous shortening is known across many of the thrust-fold belts in this region.

6.2 RECONSTRUCTION IN TERMS OF DYNAMICS

In contrast to the above more rigorous kinematic approach, in this subchapter I outline the geodynamic model of the Carpatho-Pannonian region in terms of basin evolution.

It is well documented that the last major phase of compression along the Alpine-Carpathian thrust-fold belt exhibits a clear temporal shift from the W towards the E (Jiricek, 1979). During the Pliocene only the Eastern Carpathian segment of the Carpathian loop was tectonically active (e.g. Osyczko and Slaczka, 1985). The still active subduction here is also indicated by the Pliocene volcanic activity in the area (Szabó et al., 1992). The Middle Miocene reconstruction (Fig. 6.4) shows why the southeastern part of the Carpathians remained active during the Pliocene. In this corner probably the arc still did not reach the edge of the European margin and the thin continental (or transitional) crust was able to maintain the subduction roll-back process. The larger extension in the southern part of the Pannonian Basin (see above) was partly caused by this differential movement. Again, the assumption in this reconstruction is that the total amount of extension across the Pannonian Basin definitely exceeds a value of 100 km in an E-W direction (cf. Royden et al., 1983b). In the northern part of the back-arc region the numerous calc-alkaline volcanic centers indicate the earlier onset of subduction. The thrusting gradually stopped from W to E as the West Carpathian thrust front obliquely collided with the thick continental edge of the European plate.

The Early Miocene reconstruction (Fig. 6.5) is based on the assumption that the Early Miocene volcanics along the Mid-Hungarian Line formed in association with major right-lateral movements. At the same time more distributed left-lateral shear occurred
Fig. 6.4. Reconstruction of the Pannonian/Carpithian region for the Middle Miocene.
Fig. 6.5. Reconstruction of the Pannonian/Carpathian region for the Early Miocene.
Fig. 6.6. Reconstruction of the Pannonian/Carpethian region for the Middle Oligocene.
Fig. 6.7. Reconstruction of the Pannonian/Carpathian region for the Late Eocene.
along the Pieniny Klippen Belt to the N. At this point the question arises whether this continental escape triggered the intra-Carpathian extension? A possible scenario which connects these events is outlined as follows.

Aside from the roll-back of the subducting European plate (Royden and Burchfiel, 1989), there could be an additional major driving force for the back-arc extension. As a working hypothesis I accept a model of extensional collapse (Dewey, 1988; Molnar and Lyon-Caen, 1988) in which the radial extension of the intra-Carpathian area is largely the consequence of the extensional collapse of an overthickened and therefore gravitationally unstable lithosphere (cf. Horváth and Bercikhemer, 1982). Applying this model to the Pannonian Basin and the Carpathians, I speculate that the overthickened stack of the Cretaceous-Paleogene Alpine nappes collapsed after an Early Miocene episode of eastward directed, large-scale continental escape (Kázmér and Kovács, 1985) or extrusion (Ratschbacher et al., 1991). During the Middle Miocene, radial extension and basin subsidence reflect the collapse of the elevated area of the former compressional arc.

The retroarc flexural origin of the Hungarian Paleogene Basin described in Chapter 5 and shown in Figs. 6.6 and 6.7 gives a new perspective on the Paleogene crustal-scale structure of the Carpathians. It is important to realize that while the Carpathians represented an extensional arc during the Neogene with a corresponding back-arc extensional basin (Pannonian Basin proper), the Paleogene arc of the Carpathians was more likely a compressional arc with a corresponding retroarc flexural basin (Hungarian Paleogene Basin).

The Middle Oligocene (Late Rupelian - Early Chattian) reconstruction (Fig. 6.6) assumes that the crust entering the subduction zone was the distal continental crust of the European passive margin or, else, a smaller continental ridge (continuation of the Bohemian Massif to the SE?). This could explain the sudden termination of deposition and subsequent uplift in the Lower Kiscellian (Lower Rupelian) Central Carpathian (Podhale)
forearc basin (see Late Eocene reconstruction in Fig. 6.7). The major basement-involved thrusting episode along the Hurbanovo-Diósjenő Line (Chapter 5) also corresponds to the uplift and erosion of deep basement units (e.g. Veporic) in the arc massif. This strong compressional event at the beginning of the NP24 nannoplankton zone (31 Ma or 29.5 Ma, according to Haq et al., 1987 and Harland et al., 1989, respectively) may represent the time when B-subduction along the Western Carpathians changed to A-subduction sensu Bally (1975). Along the southern part of the Carpathian arc oceanic crust was still subducting while deposition in the Szolnok and Transcarpathian forearc basins continued until the end of the Early Miocene.

Fig. 6.7 shows the Late Eocene reconstruction. During this time the Hungarian Paleogene Basin was clearly separated from the coeval Central Carpathian Paleogene or Podhale flysch basin (Samuel and Salaj, 1968). I interpret the Central Carpathian flysch basin as a forearc basin, since it was located trenchward of the continental arc and much less deformed than the outer Carpathian flysch. It is proposed here that the Central Carpathian forearc basin was separated from the accretionary complex by a ridge, i.e. the ancestral Pieniny Klippen Belt. This geometry suggests a narrow ridge-forearc setting according to the forearc classification scheme of Dickinson and Seely (1979). Since no oceanic crustal element was trapped beneath the forearc basin, it is underlain by the arc massif (various Paleozoic and Mesozoic units of the inner Western Carpathians) and the subduction complex (Magura units). This qualifies the Central Carpathian or Podhale flysch basin as a constructed forearc basin sensu Dickinson and Seely (1979).

The Magura Basin represents the trench and the northern continental slope if one accepts (for a recent discussion see Winkler and Slaczka, 1992) that indeed true oceanic crust was subducted during Late Eocene times (B-subduction). Sedimentological studies indicate only that the basin floor was at upper bathyal water depth during Late Eocene times in the Magura Trough (Ksiazkiewicz, 1975).
The interpretation of the Hungarian Paleogene Basin in terms of strike-slip tectonics related to large-scale continental escape (Fig. 1.8a) during the Early Tertiary (Middle Eocene-Late Oligocene) is essentially based on the Eocene record in the Buda Mts. The right-lateral strike-slip faults found by Fodor et al. (1992b) do not necessarily indicate the beginning of the large-scale continental escape, as was postulated by Kázmér and Kovács (1985). These strike-slip faults may simply correspond to the distributed dextral shear caused by the NW to WNW moving Adriatic plate (Fig. 6.7, cf. Platt et al., 1990).

In fact, the lower time bracket for the continental escape episode can be proven to be much younger than Middle Eocene. "Flysch" sedimentation in the presently dismembered Szolnok-Transcarpathian basin fragment (Fig. 6.1) apparently predates the large-scale right-lateral movement along the Balaton-Mid-Hungarian shear zone. In this basin the youngest sediments were deposited during the NP25 nannoplankton zone of the Late Oligocene ( Báldi-Beke et al., 1981). Similarly, in the isolated Oligocene basin fragments caught up as strike-slip duplexes to the S of Lake Balaton (see Fig. 6.1), the youngest rocks were dated as Upper Oligocene (NP25 zone, Báldi-Beke, 1984). These data constrain the beginning of the continental escape. I propose here that the escape of the North Pannonian block from the Alpine realm occurred during the Early Miocene (Fig. 6.5), in contrast to others (Kázmér and Kovács, 1985; Fodor et al., 1992), who favor a long-lasting (Middle Eocene-Late Oligocene) time period for this tectonic process.

The characteristics of the Hungarian Paleogene Basin such as the fast initial subsidence, the subsequent uplift and erosion in the individual subbasins, the migration of the depocenters, the abrupt facies changes and the fault controlled asymmetric basins, indeed meet some of the criteria given by Reading (1980) for the recognition of strike-slip basins (Báldi and Báldi-Beke, 1985). Nevertheless, all of the above mentioned criteria can also be applied to a flexural basin. In particular, the subsidence histories of pull-apart and flexural basins are not readily distinguishable (e.g. Angevine et al., 1990). Migration of
consecutive depocenters is also not a unique feature of pull-apart basins. The Middle Eocene-Early Miocene ENE-directed apparent migration of the subbasins in the Hungarian Paleogene Basin can be explained as the result of SE-directed, outward advancement of the retroarc thrust-fold belt and a superimposed NE-directed, along-strike shift of the thrust deformation. The shift along strike perhaps was caused by the oblique convergence between the Alpine - Carpathian thrust front and the Bohemian Massif (Figs. 6.6 and 6.7).

According to Royden and Báldi (1988), the broad intracontinental right-lateral transform zone responsible for the Hungarian Paleogene Basin connected the northwestern end of the Dinarides with the eastern end of the Western Carpathians (Fig. 1.8b). Recent reviews on the tectonics of the Southern Alps (e.g. Doglioni and Bosellini, 1987), however, clearly show that the NW-trending Dinaric structures do not terminate abruptly upon entering the eastern Southern Alps. Instead, they can be traced towards the NW up to the Periadriatic Line, although with gradually decreasing shortening (Doglioni, 1987). Therefore, there is no kinematic need to connect the northwestern "end" of the Early Tertiary Dinarides with the Carpathians through the intra-Carpathian region.

Regardless of the ongoing debate on the magnitude of right-lateral displacement along the Periadriatic-Balaton-Mid-Hungarian-Line system (see Chapter 8 for a detailed discussion), the Hungarian Paleogene Basin has to be palinspastically restored inside the Paleogene Alpine realm. Indeed, certain Paleogene successions in the Southern Alps have already been interpreted as being deposited in a basin controlled by S-vergent thrusts on its northern margin. The Late Eocene Ternate Formation (Bernoulli et al., 1988) and the Oligocene-Early Miocene Gonfolite Lombarda (Gunzenhauser, 1985; Gelati et al., 1988) are comparable to the sedimentary succession of the coeval Hungarian Paleogene, perhaps reflecting a similar geodynamic origin (see Fig. 6.6). Obviously the major difference is that in the case of the Western Alps continental collision took place during the Paleogene,
the case of the Western Alps continental collision took place during the Paleogene, while
in the Carpathians A-type subduction had just started.

Another interesting analogy can be drawn between the Oligocene Insbruc and the
Hurbanovo-Diójenő lines if my interpretation of the latter is correct. Heitzmann (1987)
and Schmid et al. (1987) have shown that the steeply N-ward dipping Insbruc Line is the
result of a major Middle-Late Oligocene backthrusting event in the Alps. During this
movement, rocks were uplifted to the surface from at least 10 km depth in the hanging
wall (Giger and Hurford, 1989). Similarly, in the case of the Diójenő Line, the crystalline
mass of the Veporides in the hanging wall was juxtaposed with the unmetamorphosed
Mesozoic of the Transdanubian Central Range in the footwall. Since the Veporides is in a
Middle Austroalpine structural position and the Transdanubian Central Range is in an
"Uppermost Austroalpine" position (e.g. Tollmann, 1989), the vertical component of
backthrusting should be some kilometers; however, part of this vertical offset might be the
result of earlier Eoalpine compression. At any rate, another important difference is that the
backthrusting along the Insbruc Line is dated by radiometric methods (e.g. Hurford,
1986), while the backthrusting along the Diójenő Line is dated by biostratigraphy
(boundary of NP23/24 nannoplankton zones).

As Bernoulli et al. (1990) pointed out, it is difficult to date Early Tertiary
deformations in the Southern Alps, since stratigraphic control is possible only in the
border regions. In contrast, however, Early Tertiary structural episodes can be directly
dated within the fine biostratigraphic resolution available for the Hungarian Paleogene
Basin.

The cross-sections shown in Figs. 6.6 and 6.7 also illustrate the orogenic float (sensu
Oldow et al., 1989) of the Paleogene Carpathians (A.W. Bally, pers. comm.) as a
compressional system. The major detachment surface in lower crustal depth must have
connected the Outer Carpathian folded belt with that of the Hungarian Paleogene Basin.
6.3 COMMENTS ON EOALPINE RECONSTRUCTIONS

The reconstruction of the Eoalpine position of certain tectonic provinces is not possible in terms of kinematics, since the amount of Cretaceous shortening is very poorly constrained in the Alpine system. In the following, however, I will comment on the most probable Cretaceous reconstructions in terms of dynamics. These include the Senonian "Gosau" and the Albian strike-slip tectonics in the NW Pannonian Basin, the Eastern and Southern Alps (see Fig. 5.2).

6.3.1 SENONIAN "GOSAU" BASINS

I interpreted the Senonian basin of the Transdanubian Central Range as a flexural basin in subchapter 5.4.3.1. The kinematic reconstruction of the Miocene extension brings the Senonian Basin very close (i.e. <50 km) to the Styrian (or Centralalpine) Senonian basins (Fig. 6.2). Thus the compressional origin of the Transdanubian Senonian Basin appears to be at odds with recently proposed, extensional models of Gosau basins of the Eastern Alps (Krohe, 1987; Neubauer, 1988; Ratschbacher et al., 1989; Fritz et al., 1991).

If the Gosau basins have indeed an extensional origin the conflict can be solved (Tari, 1992b) supposing that while the orogenic interior underwent extension (Centralalpine Gosau basins) shortening took place contemporaneously in the external thrust belt, resulting in a flexural basin (Transdanubian Senonian basin).

Recent papers, however, seem to favor the compressional origin of at least the Gosau basins of the Northern Calcareous Alps (e.g. Faupl and Wagreich, 1992; Wagreich, 1993). Therefore I think a geodynamic model in which the Centralalpine Gosau basins (Kainach and Krappfeld, see Fig. 5.1) are connected with the Transdanubian Senonian Basin in a retroarc flexural setting is more likely. This geodynamic model would predict a crustal-scale geometry rather similar to the Paleogene reconstruction shown in Fig. 6.7.

The double vergence of the Alps during the Senonian is well established considering
the Late Cretaceous flysch trough of the Southern Alps (Castellarin, 1976, 1977; Massari and Medizza, 1976). In fact the two-step kinematic reconstruction described in subchapter 6.1 brings the Transdanubian Central Range to the northern neighborhood of the Lombardian flysch along the Periadriatic Line (cf. Haas, 1985), suggesting their physical connection during Senonian times.

6.3.2 MIDDLE CRETACEOUS STRIKE-SLIP (TEAR?) FAULTS

The reconstruction of the minimum of 70 km NE-SW and 40 km of NW-SE extension brings close (within 100 km) the Albian dextral faults of the Bakony Mts. (Fig. 5.28) to those of the Northern Calcareous Alps. Since these right-lateral strike-slip faults share the same characteristics (Tari and Horváth, 1992), they were probably connected during Middle Cretaceous times. In fact, some of the major faults in both systems could be correlated with each other (Tari and Linzer, in prep).

Thus I postulate a continuous Upper Austroalpine cover sequence between the Northern Calcareous Alps and the Bakony Mts. as the result of large-scale Eoalpine overthrusting. This cover was eroded down to the crystalline rocks of the Middle Austroalpine during the Late Cretaceous when the Central Alpine zone uplifted. Note that this interpretation explains the problem of "exotic" Southern Alpine clasts in the Centralalpine Gosau basins (Gollner et al., 1987) which led Tollmann (1987) to postulate the hypothetic "Ultrastyrian" nappe system. In fact this enigmatic Ultrastyrian nappe did not disappear completely, but it is simply identified in this work as the Eoalpine nappe system of the Bakony Mts.
CHAPTER 7
COMPARISON WITH SIMILAR BASINS/THRUST-FOLD BELTS

This chapter attempts to compare the Neogene Pannonian Basin and the surrounding thrust fold belts to analogous systems of the world. Such a comparison highlights those general characteristics of the Pannonian Basin that may serve as a refined basis for the description of "Pannonian-type" basins (Bally and Snelson, 1980). Possible modern analogues include the Aegean-Hellenides, Alboran-Betics, Tyrrhenian-Appennines systems, the Caribbean and Indonesian regions, of course, with important differences. One ancient example will be discussed below in some detail, the Paleozoic Ouachita-Gulf of Mexico system.

Similarly, the Paleogene and Senonian basins can be also compared to other flexural basins and thrust-fold belts. A possible modern analogue for these compressional basins is provided by the Taranaki Basin, whereas the Paleogene Aquitaine-Pyrenees and Cretaceous Sevier systems are discussed as comparable ancient examples.

7.1 PANNONIAN-TYPE BASINS

In their basin classification scheme, Bally and Snelson (1980) differentiated back-arc basins associated with continental collision and on the concave side of A-subduction arc. Within this class of basins the above authors further differentiated Pannonian-type basins, i.e. where back-arc extension did not advance sufficiently to lead to opening of an oceanic basin, in contrast to W-Mediterranean-type basins that opened up as basins underlain by oceanic crust.

This definition of Pannonian-type basins is reinforced by the present work in the Pannonian Basin. Even in the most extended subbasins such as the Danube Basin characterized by low-angle normal faulting and metamorphic core complexes, the
extension was not large enough to reach the critical point where oceanic crust begins to form. This failure of the rifting process was due to the space problem imposed by the stiff continental European foreland. The thrusting in the Carpathians stopped at the edge of the thick continental crust. The buoyancy of this type of crust did not permit the continuation of subduction roll-back (e.g. Royden, 1993) so the whole system was turned off, including the extension in the back-arc region.

In contrast, the W-Mediterranean back-arc basins are not entirely confined by areas characterized by continental crust. Thus the Betic-Rif system continues to grow to the W into the Atlantic ocean, the Apennines still advance to the SE into the deep Eastern Mediterranean basin (see Fig. 1.1).

7.1.1 THE AEGEAN-HELLENIDES SYSTEM

The style of extension documented by reflection seismic data in the Aegean Sea (e.g. Martin, 1987; Mascle and Martin, 1990) is very similar to the Pannonian Basin. Although the present-day extensional style of the Aegean can be classified as wide-rift type sensu Buck (1990) interestingly enough an increasing number of Late Miocene metamorphic core complexes were recently reported by Lister et al. (1984), Buick (1991), Dinter and Royden (1993), Sokoutis et al. (1993).

As to the driving mechanism of extension, opinions differ significantly including gravitational collapse (e.g. Dewey, 1988) or roll-back of the subducting plate (e.g. Royden, 1993). According to Doglioni et al. (1991), however, the Aegean Sea cannot be regarded as a Mediterranean back-arc basin (cf. Horváth and Berckheler, 1982).

7.1.2 THE ALBORAN-BETIC-RIF SYSTEM

The Alboran-Betic system provides very fine examples of outcropping detachment faults (e.g. Garcia-Dueñas and Martinez-Martinez, 1988) analogous to those described in
Chapter 5. The stratigraphic omissions associated with these detachment faults is also significant, ranging up to 5 km (Garcia-Dueñas et al., 1992).

The driving mechanism of extension is thought to be the Miocene extensional collapse of a Paleogene "collisional ridge" (Platt and Vissers, 1989) similar to the scenario I proposed in Chapter 6 for the Pannonian Basin. Morley (1993) and Royden (1993), however, favor subduction roll-back in the Alboran-Betic system.

7.1.3 THE CARIBBEAN SYSTEM

The analogy between the Caribbean and the Pannonian Basin was first proposed by Burke and Sengör (1986) in the context of continental escape. As it is clear from the preceding chapter, escape tectonics operated before the onset of back-arc extension in the Pannonian Basin proper. Therefore this analogy holds for the Early Miocene period in the evolution of the Carpathian/Pannonian system.

Recently Mann and Burke (1990) made a point on the possibly analogous evolution of the stress-fields in the two systems as reflected by the microtectonic data (cf. Bergerat, 1989; Csontos et al., 1991).

7.1.4 THE PALEozoIC OUACHITA-GULF OF MEXICO SYSTEM

The Late Paleozoic Ouachita orogenic belt is a N-vergent thrust-fold belt, related to an A-type subduction of the Paleozoic passive margin of North America. The thin-skinned Ouachitas and their foredeep basins were formed during this southward directed subduction. Synchronously with compression in the thrust-fold belt, thick Pennsylvanian to Permian marine sediments were deposited to the S of the Ouachita belt, in the Paleozoic Gulf of Mexico. Although this sedimentary succession typically lacks apparent extensional features, its position on the concave side of an orogene associated with A-subduction points to its back-arc origin. Therefore the Neogene Pannonian Basin provides a possible
analog for this setting (Milliken, 1988, 1990; Royden et al., 1990; Tari et al., 1991).

To compare these systems two maps were compiled based on several sources showing the main structural features in both systems at the same scale (Figs. 7.1 and 7.2). Certain generalizations had to be made to arrive at a compatible legend for both maps. Note that the thrust-fold belts are directly comparable, but the back-arc basins themselves differ in their documentation. The back-arc basin of the Ouachitas is known only from several wells of the Arklatex region (Milliken, 1988) and is largely covered by the Mesozoic-Tertiary sedimentary fill of the Gulf of Mexico.

Besides the map-view similarity, the cross-sectional expression of the two systems is also comparable. A pair of crustal sketch sections was constructed (Fig 7.3) in analogous position shown in Figs. 7.1 and 7.2. Several specific details are comparable both in map and in cross-sectional view. The foreland basement promontories are of comparable size and kinematic role, such as the Llano Uplift/Bohemian Massif and the Arbuckle Uplift/Holy Cross Mts. The Arkoma Basin finds its counterpart in the Polish molasse basin. The strongly deformed zone of the Maumelle chaotic zone is directly comparable to the Pieniny Klippen Belt (Fig. 7.3). Note the similarity of S-vergent backthrusts just to the S. Farther to the S, the Benton Uplift is comparable to the Tatra Mts.

The European platform is shown with a steep southerly dip in Fig. 7.3 following the original interpretation of Tomek et al. (1989). An alternative interpretation would be similar to the one shown in the Ouachita section, i.e. the European foreland underlies the Carpathians and the Pannonian Basin much farther to the S, in lower crustal depths (cf. regional transect presented in Chapter 5, Plate 10).

In general this comparison appears to be "symmetric" between the folded belts, since both the Carpathians and the Ouachitas are classical sites of geological research. This balance holds even if these folded belts were formed with a time difference of some 300 Ma. Regarding the corresponding back-arc basins the Pannonian Basin is far better
Fig. 7.2. Simplified geology of the Neogene Carpathians - Pannonian Basin system.
Fig. 7.3. Simplified crustal cross-sections through the Ouachitas and the Western Carpathians. Locations are indicated on Figs. 7.1 and 7.2, respectively.
known, since this basin complex represents the latest basin in the evolution of that area. In the Gulf of Mexico, however, an extremely thick Mesozoic-Tertiary sedimentary blanket hides the Late Paleozoic back-arc basin (Milliken, 1988).

7.1.5 METAMORPHIC CORE COMPLEXES OF THE WESTERN UNITED STATES

At this point it seems worthwhile to point out a specific structural analogy between the Colorado River extensional corridor of Arizona and California (Howard and John, 1987; Spencer and Reynolds, 1991) and the structures of the Rába River extensional corridor of Austria and Hungary as defined in this thesis.

In contrast to many "classical" core complexes of the western United States, the footwall of the Rechnitz window is not characterized by regional scale mylonitic fabric (Pahr, 1980; Ratschbacher et al., 1990). This implies that deformation occurred dominantly in the brittle, seismogenic regime. Depending on the original thermal gradient this would imply middle crustal depths (10-15 km). Similar deformational style was observed in the Chemehuevi Mountains of California (John, 1987; John and Foster, 1993).

In fact the Chemehuevi Mountains provide other analogue aspects with the Rechnitz area. John (1987) reported the propagation of faulting into hangingwall of consecutive detachment faults such as the Mohave Wash, Chemehuevi and Devils Elbow faults. As to the relative timing same relationship was observed in the upper plate of the Rechnitz system, namely in the case of the Ikva, Répce and Rába detachment faults. This deformational sequence is regarded as characteristic for low-angle normal fault systems, according to John (1987). An additional structural feature observed in both the Chemehuevi and Rechnitz areas is the map view corrugation of the detachment faults parallel to the slip-direction with wavelengths of 10-15 km (see Fig. 5.15b). The sedimentary record of tectonic denudation such as clast type and size changes is also comparable (e.g. Miller and John, 1988 and Tari, 1993).
As to the crustal scale characteristics the geometry of the Moho should be mentioned. McCarthy et al. (1991) observed a flat Moho under the Whipple Mts. which requires the flow of lower and middle crustal material under the metamorphic core complex (see Block and Royden, 1990). McCarthy et al. (1991) were indeed able to document the thickening of the midcrustal layer beneath the Whipple Mts. The Moho also remains flat in the direction of transport in the Rechnitz system (subchapter 5.4.1.3.2). There is no evidence at present for the thickening of lower crustal material under the Rechnitz area, although I postulated it (Fig. 5.16).

7.2 PERISUTURAL BASINS

In their basin classification scheme, Bally and Snelson (1980) differentiated perisutural basins associated with continental collision. This type of basin can also be found in the intra-Carpathian area during the Mesoalpine and Eoalpine periods. Their well-known potential analogs include the basins listed below.

The Taranaki Basin in New Zealand (e.g. Stern and Davey, 1990) and the Aquitaine Basin in France (e.g. Choukroune et al., 1989) may serve as the analogues for the Hungarian Paleogene Basin. The Late Cretaceous Sevier orogenic belt and the associated foredeep basin in the western United States (e.g. Wells et al., 1990) may provide a well-studied analogue for the Transdanubian Senonian Basin as I suggested recently (Tari, 1992b).
CHAPTER 8
DISCUSSION

This chapter returns to the regional tectonic problems of the Pannonian Basin and the surrounding thrust-fold belts. The discussion here is focused on several contradicting aspects of the Alpine evolution of eastern Central Europe. My speculative models are offered here as alternatives to existing ones proposed by others and they should be rigorously tested in the future.

8.1 NEOALPINE EVOLUTION

As I mentioned in Chapter 2, the temporal subdivision of the Alpine evolution in this work is slightly different from that of Trümpy (1973). In a "Pannonian" context it seems more logical to put the boundary between the Mesoalpine and Neoalpine stages in the Early Miocene rather than in the Middle Oligocene (cf. also Trümpy, 1980). Thus the Miocene-Recent strike-slip dominated "escape" period and the subsequent back-arc extension (Neoalpine) are clearly separated from the previous Eocene-Oligocene compressional period (Mesoalpine).

8.1.1 QUATERNARY-RECENT TECTONICS

The interpretation of neotectonic features in the Pannonian Basin in terms of compressional inversion is one of the original results of this thesis. This compression seems to propagate into the Pannonian Basin from the west (see e.g. Fig. 3.40). Indeed, in-situ stress measurements showed compressive stresses in the western part of the basin, while tensile stresses were obtained in the eastern part (e.g. Dövényi and Horváth, 1990). It is not clear, however, that the present stress configuration can be regarded as representative of the late Neogene as well. Moreover, the initiation of the late-stage
compression is not very well constrained in the whole basin although it is relatively well known in some areas (Fig. 4.21). Whether the onset of inversion was the same everywhere, indicating large-scale intra-plate stress changes (i.e. between the European and African plates) or whether it commenced in a time-progressive manner, suggesting more local causes (e.g. influence of the nearby Adriatic promontory or ascending mantle diapir), is not clear at the moment. The critical step in discriminating these alternatives is the systematic study of all neotectonic features in the surrounding folded belts, including their more accurate age dating.

8.1.2 LATE MIOCENE-PLIOcene POST-RIFT TECTONICS

An important problem related to post-rift tectonics is the analysis of thermal subsidence (e.g. Sclater et al., 1980; Royden et al., 1983a,b; Royden and Dövényi, 1988). The chronostratigraphic resolution used in these subsidence calculations within the very thick post-rift sedimentary fill was very poor. The same holds for the determination of paleowater depth. Although the overall results of the above mentioned subsidence studies are probably not sensitive to these parameters, recent progress in the chronostratigraphic subdivision of the post-rift fill of the Pannonian Basin using sequence stratigraphy (Tari et al., 1992a,b; Vakarcz et al., 1992, in press) offers much more accurate input parameters for future subsidence studies.

As to the syn-rift/post-rift boundary, the regional sections of Chapter 3 provided a strong argument that the Sarmatian succession belongs to the post-rift phase, in contrast to the commonly held opinion that this boundary is located at the Pannonian/Sarmatian boundary (see Fig. 3.4). In my view the top of the syn-rift phase can be as old as Late Badenian in some places, assuming that this boundary may be diachronous throughout the basin. This preliminary proposition could be tested in Northern Hungary where the critical Badenian-Sarmatian succession has numerous outcrops and has a very well defined
lithostratigraphic framework (e.g. Hámor, 1985).

In my opinion the widespread Late Miocene-Pliocene alkaline basalt volcanism (e.g. Szabó et al., 1992) is not sufficiently understood. Several models were proposed (e.g. Embey-Isztin et al., 1990), but all of them are descriptive and rather general. The peculiar timing of this volcanism, following the syn-rift tectonism and related calc-alkaline volcanism with a time gap of about 2-5 Ma, should be a key element in any geodynamic model seeking the explanation. Moreover, the cessation of this volcanism about 2 Ma ago roughly coincides with the onset of late-stage compression and inversion discussed above.

8.1.3 MIDDLE MIOCENE SYN-RIFT TECTONICS

Regarding the kinematics of syn-rift tectonics, the magnitude of extension in the subbasins of the Pannonian basin system is generally not constrained by structural studies but only estimated by subsidence and thermal data (e.g. Royden and Dövényi, 1988). Since the wells used in these studies were mainly located in basement highs, the amount of crustal extension is probably in some cases underestimated (cf. Sawyer, 1986). Calculation of the extension from the geometry of normal faults observed on seismic reflection sections also underestimates the amount of upper crustal extension (e.g. Sclater and Shorey, 1989; Sclater and Célérier, 1989). An alternative approach to estimating the extension values would be systematic mapping of offset structural markers in the pre-extensional basement by reflection seismic data. This rigorous method could have been tried in the NW Pannonian Basin where I mapped Eoalpine nappes within the basement. However, in the Great Hungarian Plain this type of detailed work still needs to be done when the improvement of standard exploration seismic data will allow it. At any rate, the total amount of syn-rift extension across the Pannonian Basin is probably about 200-250 km in an E-W direction (see Fig. 6.1), or at least twice as much as suggested by Royden et al. (1983b), Royden (1988).
A much debated problem concerning the syn-rift stage of the Pannonian Basin is the driving mechanism for the extension (for a recent overview see Horváth, 1993). The suggested models can be subdivided into four groups.

a) Subduction roll-back or slab pull. Following Dewey (1980), this mechanism was proposed by many (Royden et al., 1982; Royden et al., 1983a,b; Royden, 1988; Royden and Burchfiel, 1989; Csontos et al., 1992). In these models the extension is driven by the negative buoyancy of the subducting slab (e.g. Royden, 1993). As to kinematics Doglioni et al. (1991) and Doglioni (1992) also proposed a model of back-arc extension driven by subduction roll-back to explain the formation of the Pannonian Basin. However, this roll-back of the European continental margin, according to Doglioni (1990), is the consequence of an eastward directed mantle flow exerting a push on the westward-dipping subducting slab. This type of dynamic explanation, however, fails to explain the formation of a number of other back-arc basins, most notably the Aegean Sea, as correctly pointed out by Royden (1993). Morley (1993) pointed out that shallow emplacement of the astenosphere under the basin is also necessary to explain the large thermal sag basin superimposed on the syn-rift basin.

b) Gravitational collapse. This mechanism, following Le Pichon and Angelier (1979) and Horváth and Berckhemer (1982), was generalized by Dewey (1988) and formulated by Molnar and Lyon-Caen (1988). Extensional collapse means extension driven by differential stresses imposed by gravity on a topographic high. The radial nature of extension in the intra-Carpathian area seems to support this interpretation (Horváth and Berckhemer, 1982).

c) Synchronous tectonic escape and extensional collapse or extrusion. Extrusion tectonics is defined as a synchronous interaction between tectonic escape (Burke and Sengör, 1986) and extensional collapse (Dewey, 1988). In a recent synthesis on the structure of the Eastern Alps, Ratschbacher et al. (1991) suggested extrusion tectonics as
the driving mechanism for the Neogene extension in the Eastern Alps and the intra-Carpathian region. Ratschbacher et al. (1991) also concluded that the forces applied to the boundaries of the Eastern Alps, causing their continental escape, seem to be of similar importance as the extensional spreading of gravitationally unstable crust.

d) Diachronous extensional collapse and subduction roll-back. As a working hypothesis the model of extensional collapse (Dewey, 1988, Molnar and Lyon-Caen, 1988) was accepted by Tari et al. (1992b), in which the initial extension of the intra-Carpathian area is the consequence of the extensional collapse of an overthickened and therefore gravitationally unstable lithosphere (cf. Horváth and Berckhemer, 1982).

Applying this model to the Pannonian basin and the Carpathians, one can speculate that the Early Miocene episode of eastward directed, large-scale continental escape described in Chapter 6 was responsible for the disintegration of the former Cretaceous-Paleogene compressive arc. Once the compressive boundary conditions were removed, the overthickened stack of Alpine nappes began to collapse. During the Middle Miocene, radial syn-rift extension reflected the collapse of this elevated area. The extension was accommodated by coeval shortening in the form of radial thrusting on the European continental margin, forming the outer, Tertiary units of the Carpathians. This collapse mechanism was no more effective after the remarkable thickness and elevation contrast between the Alpine orogenic wedge and the European foreland largely diminished. However, once the back-arc extension was triggered by the collapse, the slab-pull mechanism could take over and continue to drive the back-arc extension in a self-sustaining manner.

It is well documented that the last major phase of compression along the Alpine-Carpathian thrust front exhibits a clear temporal shift from the west towards the east (Jiricek, 1979; Oszczypko and Slaczka, 1985). The main direction of coeval extension in the Pannonian Basin was subnormal to this curvilinear zone of compression (Csontos et
al., 1991). The retreating of the European slab is also manifested by the development of a volcanic arc 100 to 150 km behind the front of underthrusting (e.g. Szabó et al., 1992).

The temporal evolution of the roll-back mechanism was most probably determined by the buoyancy of the subducting slab. Once the thick continental crust of the European plate entered the subduction zone, the whole process must have been terminated, giving way to molasse sedimentation in the foredeep. Thus the edge of the thick continental crust of the European plate imposed the most important boundary condition for the evolution of the Tertiary Carpathian thrust-fold belt and the associated Pannonian back-arc basin (see reconstructions of Chapter 6).

Alternatively, the process of lithospheric downbending could have been terminated, eventually, by the detachment of the subducted slab. Seismic tomography and subcrustal seismicity can detect a remnant slab only at the southernmost corner of the Eastern Carpathians (Oncescu, 1984a,b). Tari et al. (1992b) speculated that the subducted slab has been detached progressively from west to east along the Carpathian arc. Slab detachment leads to termination of foredeep subsidence, uplift, and change in the stress regime of the back-arc region as proposed for the Aegean area by Sorel et al. (1988).

8.1.4 EARLY MIOCENE ESCAPE TECTONICS

Perhaps the most critical step in the reconstruction is the restoration of the Northern Pannonian unit in the Alpine realm. Géczy (1973) was the first to propose large-scale strike-slip movements to explain the present-day "erratic" relation of the Northern Pannonian unit (with southern, "African" affinity) to the Southern Pannonian, Tisia unit (with northern, "European" affinity). Since the continental escape (sensu Burke and Sengör, 1986) model of Kázmér (1984) it is generally accepted that the Northern Pannonian unit was squeezed out to the E from the Alps during the Tertiary (e.g. Kázmér and Kovács, 1985; Balla, 1985, 1988; Neubauer, 1988; Ratschbacher et al., 1991; Schmidt
et al., 1991; Csontos et al., 1992). This major eastward directed movement involved large-scale strike-slip faulting along the southern boundary of the escaping unit which is outlined by the composite Periadriatic-Balaton-Mid-Hungarian Lines (Fig. 8.1) by all the above authors. However, the suggested 400-500 km of right-lateral offset along the eastern end of the Periadriatic Line obtained from the correlation of offset Permo-Mesozoic facies zones (e.g. Majoros, 1980; Kázmér and Kovács, 1985) seems to be unrealistic for Alpine geologists (e.g. Carlo Doglioni, pers. comm.). Indeed, much smaller offsets can be observed at the western end of the Periadriatic Line (e.g. Laubscher, 1991).

To clarify this controversy I compiled all the available estimates on the magnitude of right-lateral offset along the entire length of the Periadriatic Line (Fig. 8.2). These include estimates based on displaced facies boundaries and structural markers. Ideally these markers should be well-defined vertical boundaries striking perpendicular to the Periadriatic Line. However, these markers are typically not well-defined and/or strike obliquely to the strike-slip zone, resulting in an error bar of the offset estimation. Therefore I indicated an estimated error for the markers to show their relative reliability.

The oldest facies marker is provided by the particular geometry of the pre-Alpine (Hercynian) metamorphic zones in the Southern Alps. These zones trend perpendicular to the Periadriatic Line, displaying a gradually decreasing grade of metamorphism to the E (Vai and Cocozza, 1986). Árkai (1987) observed a similar eastward decreasing tendency in the metamorphic grade of the Transdanubian Central Range. Here I tentatively correlate the facies boundary between the medium- and low-grade facies Hercynian basement with a large error bar.

Recently Ebner et al. (1991) pointed out the similarity between the Carboniferous facies of the Carnian Alps-South Karawanken (Italy, Austria, Slovenia) and the Upony-Szendrő Mts (Hungary). This facies boundary, however, has a poorly known strike and therefore does not give a good constraint on the offset.
Fig. 8.1. Map of facies and structural markers along the Periadriatic Line. See also Fig. 8.2 on next page.
<table>
<thead>
<tr>
<th>Marker</th>
<th>Author(s)</th>
<th>Range(km)</th>
<th>Offset</th>
<th>Error(±)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Facies markers:</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hercynian metamorphics</td>
<td>this work</td>
<td>100-640</td>
<td>540</td>
<td>50</td>
</tr>
<tr>
<td>Carboniferous clastics</td>
<td>Ebner et al. (1991)</td>
<td>480-1000</td>
<td>520</td>
<td>100</td>
</tr>
<tr>
<td>Upper Permian sediments</td>
<td>Majoros (1980)</td>
<td>300-800</td>
<td>500</td>
<td>30</td>
</tr>
<tr>
<td>Middle Triassic volcanites</td>
<td>Horváth and Tari (1987)</td>
<td>370-780</td>
<td>410</td>
<td>50</td>
</tr>
<tr>
<td>Upper Triassic carbonates</td>
<td>Kázmér and Kovács (1985)</td>
<td>350-800</td>
<td>450</td>
<td>30</td>
</tr>
<tr>
<td>Lower Jurassic carbonates</td>
<td>Kázmér (1987a)</td>
<td>170-670</td>
<td>500</td>
<td>30</td>
</tr>
<tr>
<td>Jurassic ophiolites</td>
<td>Laubscher (1971)</td>
<td>560-1040</td>
<td>480</td>
<td>50</td>
</tr>
<tr>
<td>Lower Cretaceous clastics</td>
<td>Kázmér (1987b)</td>
<td>170-710</td>
<td>540</td>
<td>30</td>
</tr>
<tr>
<td>Bergell-Gonfolite</td>
<td>Heitzmann (1987)</td>
<td>30-70</td>
<td>60</td>
<td>30</td>
</tr>
<tr>
<td>Smrekovce-Bugyi volcanoes</td>
<td>Nagymarosy (1990a)</td>
<td>520-860</td>
<td>340</td>
<td>50</td>
</tr>
<tr>
<td>Oligocene basins</td>
<td>Nagymarosy (1990a)</td>
<td>580-1020</td>
<td>440</td>
<td>50</td>
</tr>
<tr>
<td>Late Oligocene clastics</td>
<td>this work</td>
<td>400-860</td>
<td>460</td>
<td>50</td>
</tr>
<tr>
<td>Ivrea-Pejo</td>
<td>Laubscher (1991)</td>
<td>0-160</td>
<td>160</td>
<td>10</td>
</tr>
<tr>
<td>Ivrea-Mihályi</td>
<td>Balla (1992a)</td>
<td>0-560</td>
<td>&gt;400</td>
<td>?</td>
</tr>
<tr>
<td>Late Eocene Dinaric front</td>
<td>this work</td>
<td>250-650</td>
<td>400</td>
<td>50</td>
</tr>
</tbody>
</table>

**Structural markers:**

- Ivrea-Mihályi: Balla (1992a), 0-560, >400 km.
- Late Eocene Dinaric front: this work, 250-650, 400 km.

**Structural-related additional displacements included in the above apparent offsets:**

- a) E-W shortening along the Giudicaria Belt, 30-50 km (Laubscher, 1988, 1990a).
- b) E-W shortening associated with the Dinaric front in the Dolomites (Doglioni, 1987).
- c) NE-trending Late Miocene sinistral offset of the Periadriatic Line by the Giudicaria segment, 80 km (Balla, 1986; Laubscher, 1991).
- e) E-W Miocene extension in the NW Pannonian Basin, 70-100 km (this work).

Fig. 8.2. Apparent offset of structural and facies markers along the Periadriatic Line in kilometers. I applied a kilometer scale along the Periadriatic-Mid-Hungarian Line starting from the Ivrea zone of the western Southern Alps (see Fig. 8.1). The lower value in the "range" column in this table shows the projection of a given facies/structural marker to the Line on its southern side, whereas the upper value corresponds to the position of the same marker on the northern side of the Periadriatic-Mid-Hungarian Line. Their difference gives the amount of dextral offset. The error bar was estimated by me and it is intended to give only a feeling for the "accuracy" of the offset estimate. The values listed as structure-related additions should be subtracted from the apparent offsets, which significantly decrease the amount of Early Miocene offset close to the 160 km of Laubscher (1991)!
The facies boundary between the continental and shallow marine Permian rocks strikes perpendicular to the Periadriatic and Balaton ("Mid-Hungarian") Lines, both in the eastern Southern Alps (Buggisch, 1978) and in the Transdanubian Central Range as well (Majoros, 1980). The latter author was actually the first who estimated the amount of right-lateral offset along the Periadriatic-Mid-Hungarian Line, namely 500 km.

The presence of Middle Triassic volcanites besides the widespread tuff horizons is a relatively new finding in the area of the Transdanubian Central Range (Raincsák, 1980; Cros and Szabó, 1984; Kubovics et al., 1990). In the eastern Southern Alps these volcanites are quite common, and in a recent review Castellarin et al. (1988) drew a facies boundary between the acidic/intermediate and mafic suites. Based on the results of Horváth and Tari (1987) a much less defined boundary can also be drawn to the E of the Buda Mts. Peculiar rock types, such as the Rio Freddo ignimbrite outcropping to the S of Tarvisio, seems to correlate one for one with the Budaörs ignimbrite exposed just to the W of Budapest.

The very thoroughly studied Norian facies patterns (e.g. Prey, 1980; Kovács, 1982) provided the strongest argument for the continental escape model (Kovács and Kázmér, 1985; see Fig. 1.7). The Alpine-type Triassic also has a characteristic facies boundary to the Dinaric-type Triassic to the E of Budapest (Fig. 2.19a).

The Lower Jurassic (Hettangian-Sinemurian) succession of the Transdanubian Central Range finds its counterpart facies belts in the Southern Alps (Kázmér, 1987a). These belts strike perpendicular to the Periadriatic-Mid-Hungarian Line. Kázmér (1987a) suggested a correlation scheme in which the Zala Basin corresponds to the Lombard Basin, the Trento Plateau to the Bakony Mts and the Belluno trough to the Vértes-Gerecse Mts.

Laubscher (1971) was the first who assumed large-scale right-lateral movements along the Periadriatic Line, based on the correlation of Vardar ophiolites in the Dinarides
with the Meliata ophiolites in the Western Carpathians. However, since he supposed the straight continuation of the Periadriatic Line to the southern Pannonian Basin, his estimate (300 km) is about 150 km less than other estimates.

The Lower Cretaceous sedimentary facies of the Transdanubian Central Range shows essentially the same pattern as the Lower Jurassic (Kázmér, 1987b). These are again directly comparable to their counterparts in the Southern Alps.

The clasts of the Middle Oligocene Bergell pluton can be found in the Upper Oligocene to Lower Miocene Gonfolite Lombarda, suggesting a poorly constrained 60 km of right-lateral offset (Heitzmann, 1987). Laubscher (1991) argues, however, that this is a poor marker, since movements on the Periadriatic Line might have been partly synchronous with the deposition of the Gonfolite. Therefore this estimate should be regarded as a lower minimum.

The Middle Oligocene Smrekovec and Bugyi/Sári volcanic centers in Slovenia and in Hungary were tentatively correlated by Nagymarosy (1990a). The same author also lined up convincing evidence for the correlation of the Paleogene basins of Slovenia and Hungary. These presently dismembered basins once formed a single basin.

Similarly, the Upper Oligocene to Lowermost Miocene glauconitic sandstones in the easternmost Southern Alps (Massari et al., 1986) may be correlated with similar rocks in Northern Hungary (Sztanó and Tari, 1993).

The thick Lower Miocene volcanic succession outcropping to the S of the Bükk Mts. was correlated with similar rocks known from the northern foreland of the Mecsek Mts. by Csontos et al. (1992). However, these rocks have a very widespread distribution in the subsurface, and their formation must have overlapped at least partly the dextral slip along the Periadriatic-Mid-Hungarian Line. Therefore this facies marker is not taken into account in this compilation.

Unfortunately there are only a few structural markers known to give a reasonable
constraint on the offset along the Periadriatic Line. Laubscher (1991) mentioned the offset Austroalpine front and Ivrea zones, both indicating dextral slip of 160 km. While it is clear that the Ivrea zone should be correlated with the Pejo area just to the W of the Giudicaria zone (Laubscher, 1991), Balla (1992a) proposed a structural correlation with the Mihályi area of W-Hungary based on a rather ambiguous magnetic signature. Although Balla (1992a) mentioned an offset of >400 km, actually such a correlation implies 560 km of dextral slip.

All in all, the estimates of right-slip based on various facies/structural markers indicate an overall eastward increasing offset along the Periadriatic-Mid-Hungarian Line. The important point I make here is that these offset values are only apparent and they should be corrected by taking into account a number of structures formed after the Early Miocene right-slip.

To the S of the Periadriatic Line the Late Miocene N-S shortening along the Milan belt (Laubscher, 1985; Schönborn, 1992) had an E-W component along the Giudicaria Belt which is estimated as about 30-50 km (Laubscher, 1990b). Obviously this value should be subtracted from the apparent offset. Moreover, the Late Miocene left-lateral Giudicaria segment offsets an originally straight Periadriatic Line (Balla, 1986; Laubscher, 1991), by about 80 km.

To the N of the Periadriatic Line the minimum of E-W directed orogene-parallel Middle Miocene extension in the Eastern Alps was estimated as about 150 km (Ratschbacher et al., 1991b). Farther to the E the same E-W Middle Miocene extension in the Northern Pannonian domain is at least 100 km (see Chapter 6). Again, these values should be subtracted from the presently observed dextral offset.

Once these subtractions have been made a value of about 200 km can be obtained with an error bar of about 50 km along the entire length of the Periadriatic-Mid-Hungarian Line. This value is significantly lower than many of the apparent, uncorrected offsets
supporting the viability of the continental escape model. The conclusion here is that the escape model cannot be discarded just because the offsets seem to be unrealistically high.

It is worthwhile to look at the rate of strike-slip movements along the Periadriatic-Mid-Hungarian Line. In my opinion the right-slip dominantly occurred during the Early Miocene (see discussion below), between about 21 and 17 Ma (Eggenburgian-Otnangian-Lower Karpatian Paratethys stages). Accepting that the 200 km dextral slip took place during this 4 Ma time period one ends up with a slip rate of 50mm/year. This is quite acceptable, since recent estimates on the slip rate of a similar size strike-slip fault, the San Andreas fault of California vary between 30 and 40 mm/year (Wallace, 1990).

There is much confusion regarding the timing of the continental escape period. On one hand many consider the escape as exclusively Paleogene (Kázmér, 1984; Kázmér and Kovács, 1985; Fodor et al., 1992b). This opinion can be easily discarded based on the isolated Paleogene basin fragments along the Balaton Line or the termination of sedimentation in the Szolnok flysch belt, which suggests that the escape is definitely younger than Late Oligocene (see Chapter 6).

Another view is represented by Ratschbacher et al. (1991) based on the structural analysis of the Eastern Alps. Their "escape" tectonics, however, as a synchronous interaction between tectonic escape and extensional collapse can be proven to be younger than the actual large-scale displacement along the Periadriatic-Mid-Hungarian Line. Ratschbacher et al. (1991) argued that their poorly dated Miocene strike-slip fault system is the result of the indentation of the stiff Adriatic promontory into the Eastern-Central Alps. Therefore their strike-slip faults are coeval with the disintegration of the Periadriatic Line along the sinistral Giudicaria Line and the formation of the Milan belt (Middle to Late Miocene, see Laubscher, 1991). Thus even though I accept their tectonic model for the orogene-parallel extension in the Eastern Alps, this structural stage should be clearly distinguished from the earlier continental escape period.
Much less clear is the problem of the specific detachment level on which the eastward escape occurred. The movement of the whole crust or lithosphere does not seem very likely. As possible intra-crustal detachment horizons the top of Penninic seems to be an obvious candidate which was supposed to be a regional décollement during the Miocene orogene-parallel extension (e.g. Neubauer and Genser, 1990; Ratschbacher et al., 1991). The regional transect through the escaping province (Plate 10), however, suggests a regional décollement below the top of the Penninic. This surface may be placed most probably within the Penninic, at lower crustal depth.

One more point should be made in conjunction with the Early Miocene escape tectonics. The eastward moving North Pannonian block drove the clockwise rotation of the South Pannonian (Tisia) block (Fig. 1.10) in the model of Balla (1984). The corresponding large (~100°) clockwise rotation in Tisia was indeed reported by Patrascu et al. (1990) and Surmont et al. (1990) from the Apuseni Mts (Fig. 8.3a). In the kinematic reconstruction of the previous chapter (Figs. 6.2 and 6.3) I rotated Tisia in a counterclockwise manner (Fig. 8.3b), which resulted in a more coherent picture regarding kinematic markers such as Paleogene basin fragments and the Vardar ophiolites. Obviously this counterclockwise rotation violates the paleomagnetic results. The Cretaceous paleomagnetic directions reported by Patrascu et al. (1990) and Surmont et al. (1990), however, have very shallow inclinations and were obtained mostly from tectonically uncorrected volcanics. It is a matter of subjective decision whether the sense of rotation is regarded as clockwise or counterclockwise. I speculate that the counterclockwise rotational scheme (Fig. 8.3) might be proven to be reasonable by future paleomagnetic measurements.

8.2 MESOALPINE EVOLUTION

The new geodynamic model for the Paleogene evolution of the Carpathians and
Fig. 8.3. Contrasting models on the rotation of the Southern Pannonian (or Tisia) unit. The traditional model (a) considers clockwise rotation (e.g. Balla, 1984), whereas in this thesis counterclockwise rotation (b) is proposed as a speculation. Although the counterclockwise rotation violates existing paleomagnetic results (see text) it has advantages such as a more coherent Early Alpine paleogeography (cf. Figs. 1.5 and 1.6).
Pannonian region contradicts several existing models. In the Polish sector of the Carpathians the Pieniny Klippen belt is interpreted as an arc in the fundamental works of Birkenmajer (1986, 1988). As Fig. 6.7 shows, the Klippen Belt was a ridge during the Early Paleogene, separating the Podhale forearc basin from the actual ridge farther to the S. The new model also challenges the existing classification of the Szolnok flysch basin as a trench (Balla, 1982). This misconception is based on the correlation with Outer Carpathian units which should be discarded based on the findings of Nagymarosy and Báldi-Beke (1993). These latter authors correlated the Szolnok flysch with the Podhale basin. My new model which physically connects the Podhale basin with the Szolnok flysch basin (Fig. 6.7) appears to solve the old debate (e.g. Szepesházy, 1973; cum. lit.) on the "flysch" character of the latter basin, suggesting its forearc origin.

An essentially unsolved problem is posed by the enigmatic Paleogene magmatism not only in the intra-Carpathian region but in the Alps as well (see Polino et al., 1990 for a discussion). This post-collisional magmatism (Dal Piaz et al., 1988) was postulated to be the result of an Oligocene extensional period or "lull" (Laubscher, 1986). Laubscher (1986) based his interpretation on the apparent Oligocene extension observed on the European foreland. Furthermore, Laubscher (1988) presents the Hungarian Paleogene Basin as the manifestation of the Oligocene extension. My detailed interpretation of the same basin in terms of compression (Chapter 5) contradicts the view of widespread Oligocene extension in the Alps.

Looking at the distribution of Eocene and Oligocene volcanic centers (Fig. 5.19), it is important to realize that they are not centered in the arc (cf. reconstructions on Figs. 6.6 and 6.7). Instead the volcanic centers are located to the S of the arc mostly within the retroarc basin. This could suggest the very shallow character of the subduction of Europe to the S. A further complication comes from the fact that similar Eocene volcanics are known from the northeastern part of the Dinarides related to the subduction of Adria to
the NE. In the palinspastic reconstruction (Fig. 6.7) these Dinaric volcanics are located close to the ones in the Hungarian Paleogene Basin, which may point to the possibility of deriving the latter from the Adriatic subduction.

8.3 EOALPINE EVOLUTION

The interpretation of the Graz Palaeozoic as a Middle Miocene extensional allochthon contradicts recently proposed models on Late Cretaceous extension in the same region. The contact of the Graz Palaeozoic with the underlying Middle Austroalpine and other tectonic contacts within the Graz Palaeozoic are documented low-angle normal faults (Krohe, 1987; Neubauer, 1989; Fritz et al., 1991). These extensional faults are postulated to be Senonian by the above authors, based on the inferred causal relationship with the adjacent Kainach Gosau basin (see Fig. 5.13a). The model of Late Cretaceous extension then was extrapolated for all of the Gosau basins of the Eastern Alps (Ratschbacher et al., 1989). Here I challenge this interpretation since, based on the Rechnitz-Wechsel metamorphic core complex model described in Chapter 5, most if not all of the low-angle normal faults at the eastern end of the Eastern Alps should be Miocene in age.

The relative structural position of the Transdanubian Central Range and the Bükk Mts. is a matter of conceptual interpretation at present. According to Csontos (1988) and Csontos and Vörös (in press) the Transdanubian Central Range is structurally above the Bükk. I argue, however, that the corollary of the Eoalpine evolutionary scheme presented in Fig. 5.33 is that the Bükk Mts. are thrust on top of the Transdanubian Central Range along a poorly defined Vardar "suture" zone.

The reconstruction of the Vardar zone s.l. from its pre-extensional configuration by the counterclockwise rotation of Tisia (Figs. 6.2 and 6.3) results in a double-vergent NW-trending Eoalpine ophiolite belt (Fig. 8.4). A comparable double-vergent ophiolite belt is known in the Phillipines (A.W. Bally, pers. comm.). The Vardar ophiolite belt formed
from the Vardar ocean which had no direct connection during Jurassic times with the
coeval Penninic ocean (e.g. Pober and Faupl, 1988). The subsequent complicated multi-
stage deformational history exemplified by the results of this thesis obscured this initial
spatial separation. This is the reason why Laubscher (1971), who assumed the existence of
only one Tethyan ocean, found the palinspastic reconstruction of the Alps and Dinarides
hardly feasible. This dilemma, known as the Alps-Dinarides problem, was declared still
unsolved by Laubscher (1988). I believe that Fig. 8.4 clearly shows an explanation for this
"problem".
Fig. 8.4. Simplified model of Mesozoic ophiolites in eastern Central Europe. The "topologic" display of 3D geometric relationships was inspired by Laubscher (1971).
CHAPTER 9
CONCLUSIONS

The new observations and interpretations of this thesis are summarized below. I try to distinguish observations from interpretations. The following conclusions are based on the original work described in Chapters 3, 5 and 6.

9.1 EARLY ALPINE (PERMIAN-EARLY CRETAEOUS) EVOLUTION

Observation:

1) Currently available seismic reflection data do not reveal any clearly Early Mesozoic structure in the basement of the Pannonian Basin.

9.2 EOALPINE (MIDDLE CRETAEOUS-PALEOCENE) EVOLUTION

Observations:

1) The Transdanubian Central Range and the basement of the Danube Basin are an allochthonous Eoalpine nappe complex.

2) The WNW-trending right-lateral strike-slip faults in the Bakony Mts. are post-Barremian but pre-Senonian in age, with subsequent reactivations.

3) The sedimentary facies of the Senonian basin reflects the underlying Eoalpine nappes due to the inherited paleotopography.

4) Intra-Senonian unconformities and breccias show the growth of reactivated thrusts; thus the Senonian basin is a syn-tectonic succession.

Interpretations:

5) An Aptian stage of compression formed WSW-vergent thin-skinned thrust sheets in the Zala Basin.

6) The first-order syncline of the Transdanubian Central Range is Albian in age and
resulted from thick-skinned thrusting beneath the Velence anticlinorium.

7) The WNW-trending Albian right-lateral strike-slip faults are tear faults and detached within the Eoalpine Upper Austroalpine nappe system.

8) The Transdanubian Senonian basin is a flexural basin in contrast to the extensional (?) Centralalpine (Styrian) Gosau basins, and it was controlled by a SE-vergent (present-day coordinates) thrust system. Since thrusting was antithetic to the subducting European plate, a retroarc flexural basin setting is therefore proposed.

9.3 MESOALPINE (EOCENE-OLIGOCENE) EVOLUTION

Observations:

1) Currently available seismic reflection data do not reveal any clearly Paleogene structures in the Hungarian Paleogene and the Szolnok flysch basins.

2) The Paleogene of the Transdanubian Central Range has an erosional boundary to the NW under the Danube Basin.

Interpretations:

3) The Middle Eocene-Early Oligocene evolution of the Hungarian Paleogene Basin shows the characteristics of the early, "flysch" stage of a flexural basin.

4) The Late Oligocene-Early Miocene evolution of the Hungarian Paleogene Basin displays the typical features of the late "molasse" stage of a flexural basin.

5) The Diósjenő-Hurbanovo Line is a basement-involved Middle to Upper Oligocene thrust comparable to the Insubric Line of the Alps.

6) The Middle Eocene-Early Miocene migration of the depocenters of the Hungarian Paleogene Basin to the ENE (present-day coordinates) is the result of the advancement of thrusting to the SE and the along-strike shift of the main deformation to the NE in the thrust belt.

7) The Hungarian Paleogene Basin was located along the inner side of the Paleogene
Carpathian arc and was controlled by a thrust system which was antithetic to the subducting European plate; therefore it represents a retroarc flexural basin complex.

8) The compressional arc of the Western Carpathians developed above a B-subduction zone during the Late Eocene with a corresponding trench (Magura), a forearc basin (Podhale) and a retroarc flexural basin (Hungarian). From the Late Oligocene to the Early Miocene the Carpathian arc was characterized by A-subduction with a corresponding peripheral foredeep (Dukla and Silesian) and retroarc flexural basin (Hungarian).

9) The restored position of the Szolnok flysch basin is correlatable with the Podhale flysch and is a forearc basin.

10) Restoring the Hungarian Paleogene basin to its original position in the Alpine edifice suggests a correlation with the Gonfolite Basin of Lombardy.

9.4 NEOALPINE (MIOCENE-RECENT) EVOLUTION

Observations:

1) The postulated Early Miocene right-lateral strike-slip along the Balaton and Mid-Hungarian Lines cannot be resolved using the currently available seismic reflection data.

2) The Rába Line is not a major subvertical strike-slip fault with large horizontal offsets but instead is a low-angle normal fault with laterally changing dip-slip offsets.

3) The Middle Miocene upper crustal syn-rift extension is unevenly distributed in the Pannonian Basin. On the one hand, some areas are not extended and, on the other hand, other areas are strongly extended (>200%).

4) These contrasting areas characterized by different amounts and/or polarity of extension appear to be separated by transfer faults.

5) In the strongly extended areas (Danube, Hód and Békés Basins) extension was accommodated by detachment faults, locally forming metamorphic core complexes (e.g.
the Rechnitz Window group in the Rába extensional corridor).

6) The detachment faults interacted with the earlier Eoalpine décollement levels.

7) There are fewer syn-rift pull-apart basins than previously thought (Kiskun, Csepel and Derecske basins).

8) The dominantly lacustrine post-rift succession may be described in terms of third- and fourth-order scale sequence stratigraphy.

9) The frequently observed Quaternary - Recent flower structures in the post-rift sedimentary succession have little to do with the formation of the Miocene Pannonian Basin; instead, they are neotectonic features.

10) The outcrop pattern in the Pannonian Basin is determined by the Quaternary - Recent arching of the pre-Tertiary basement underlying the Neogene basin occurring at short (5-30 km) and long (50-100 km) wavelengths.

*Interpretations:*

11) The disintegration of the Early Tertiary compressional arc of the Carpathians due to the continental escape of the North Pannonian block from the Alpine realm occurred in the Early Miocene (19-17.5 Ma, ~Otnagian).

12) After the relaxation of the compressive boundary forces, the North Pannonian block extensionally collapsed during the Middle Miocene (17.5-16.5 Ma, ~Karpatian).

13) The syn-rift extension can be subdivided in the NW Pannonian Basin into an earlier metamorphic core complex type and a subsequent wide-rift style rifting.

14) The syn-rift extension was enhanced by the initiation of subduction roll-back of the European plate beneath the Carpathians, becoming the main driving mechanism by the late Middle Miocene (16.5-14 Ma, ~Early to Middle Badenian).

15) Continental extension could not advance to the localization of extension into a narrow rift, except perhaps the center of the Danube Basin.

16) Widespread Upper Badenian and Sarmatian strike-slip faulting belongs to the
post-rift phase and records a basin-wide inversional stage.

17) The Quaternary - Recent basement upwarps by thrusting and/or folding and the numerous flower structures suggest the beginning of late-stage compression. This inversion propagates into the Pannonian Basin from the W, from the Dinarides - Southern Alps junction area.
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APPENDICES

APPENDIX A

PHANEROZOIC STRATIGRAPHY OF THE NEIGHBORING AREAS IN THE ALPINE-CARPATHIAN-PANNONIAN JUNCTION ZONE

The stratigraphic description in this appendix covers the regions located adjacent to the actual study area described in Chapter 4. These regions include the eastern end of the Eastern Alps with the Styrian basin (Austria), the western end of the Western Carpathians (Slovakia), and the junction area in between with the Vienna basin (Austria/Slovakia; see Fig. A.1).

A.1 PRE-ALPINE (PRE-PERMIAN) STRATIGRAPHY

Since the pre-Alpine basement of the Penninic units is not known in the transitional zone between the Alps and the Carpathians, only the Austroalpine nappe complexes are discussed in this subchapter.

A.1.1 LOWER AND MIDDLE AUSTROALPINE UNITS

The Lower Austroalpine outcrops in a broad region at the eastern end of the Eastern Alps (Fig. A.1a). It is subdivided into two major systems. The lower Wechsel system has three basement units: the Wechsel Gneiss, the Waldbach Complex and the Wechsel Phyllite. The upper Grobgneiss/Raabalpen unit has also a polymetamorphic origin.

The deepest levels in the Wechsel Window (Fig. A.1b) are characterized by chlorite-bearing gneisses (*Wechsel Gneiss*) locally intercalated with phyllites and greenschists. The gneiss sequence has a gradual transition to the overlying *Wechsel Phyllite*, which is a quartz-phylilitic with significant graphite-phylilitic and quartzite interbeddings. The *Grobgneis* unit structurally overlies the Wechsel complex in large recumbent folds. In the
Fig. A.1a. Major tectonic units of the Alps-Carpathians-Pannonian Basin junction area.
Fig. A.1b. Index map of the Alps-Carpathians-Pannonian Basin junction area.
northern part, phyllitic mica-schists dominate the lithology, while to the S quartz-rich mica-schists and amphibolites occur. The Grobgneis outcrops also in the Leitha Mts. (Fig. A.1).

The Middle Austroalpine basement dominates the central zone of the Eastern Alps to the E of the Tauern Window (Fig. 2.8). To the W of the Wechsel Window, the Middle Austroalpine is composed of distinct tectonostratigraphic units, the deeper Muriden and higher Koriden polymetamorphic complexes (e.g. Neubauer et al., 1992). To the E of the Wechsel area, the Sieggraben complex can be found in some klippen.

The Muriden complex is made up of three tectonostratigraphic units from bottom to top (Neubauer and Frisch, 1992): the "Core", the Speik and the Micaschist-Marble complexes. The "Core complex" outcrops in the Gleinalm metamorphic dome (e.g. Neubauer, 1988) and has a number of lithologic varieties: strongly foliated biotite paragneisses, orthogneisses, amphibolites, metatonalite, granite and granodiorite. The complex is regarded as a late Proterozoic to early Paleozoic root of a magmatic arc (Neubauer and Frisch, 1992). The Speik complex is made up of some hundred meters of thick garnet amphibolites and thin augengneiss or serpentinite lenses up to 20 km length. The whole succession is interpreted as an ophiolite sequence of Early Paleozoic (pre-Silurian) age (Neubauer et al., 1989). The lithology of the uppermost Micaschist-Marble complex of the Muriden unit shows the following pattern: feldspar- and quartz-rich micaschists in the lower part, garnet micaschists, quartzites in the middle part and marbles, amphibolites and pegmatites can be found in the upper part. The whole sequence can be correlated with the fossil-bearing Late Ordovician to Early Carboniferous sequences of the structurally higher units of the Eastern Alps. Thus micaschists are supposedly Late Ordovician to Early Devonian in age and the marbles are Middle Devonian in age (Neubauer, 1988).

The Koriden complex is exposed in the Koralpe, Saualpe and Pohorje Mts. (Fig.
A.1). It consists of a sequence of paragneiss with kilometer-scale eclogite bodies in it (e.g. Hinterlechner-Ravnik and Moine, 1977). Intercalations of metagabbro, marble, quartzite and pegmatite are also common. The complex is interpreted as a Hercynian metamorphic succession intruded by numerous pegmatites in late Variscan times. The eclogites are considered to be Permian or post-Permian(?) in age (Neubauer, 1991).

The **Sieggraben complex** in the klippen of the Middle Austroalpine to the E of the Wechsel area (Fig. A.1) is composed of paragneiss, eclogite amphibolite, marble and serpentinized peridotite. The sequence can be correlated with the gneiss complex of the Koriden.

The pre-Alpine basement complexes of the **Tatricum** (Lower Austroalpine) in the Little Carpathians (Fig. A.1) were subdivided recently into four major lithostratigraphic units by Plasienka et al. (1991). These are the **Pezinok, Pernek, Harmónia** and **Dolany** formations. They commonly comprise two dominant lithologies, named A and B. The lower A formation shows a predominantly siliciclastic, flysch-like lithology probably Silurian to Lower Devonian in age which has a gradual transition to the overlying volcanosedimentary B formation. The latter is composed of Middle Devonian black shales, quartzites, marls, basalts and their tuffs. Locally, gabbro bodies were also found.

There are two important Hercynian plutons in the Tatricum of the Little Carpathians, the Bratislava granite-monzonite and the Modra granodiorite bodies (Plasienka et al., 1991). Similar Tatric successions outcrop in the core of the Inovec and Tribec Mts. as well (Fig. A.1) and the same rocks were found in wells in the basement of the northwestern part of the Slovakian Danube Basin (Fusán et al., 1987).

In contrast, however, the basement of the southeastern part of the Slovakian Danube Basin and the area of Northern Hungary at the Börzsöny Mts. (Fig. A.1) are part of the **Middle Austroalpine Veporides** (Fusán et al., 1987; Fülop, 1990). These sequences comprise greenschist to amphibolite facies metamorphic rocks.
A.1.2 UPPER AUSTROALPINE UNITS

The Upper Austroalpine is distributed as large klippen in the central and southern areas of the Eastern Alps (Fig. A.1). The pre-Alpine stratigraphy of these areas is described below beginning with the Gurktal Complex, the Northern Karawanka Mts. and Pohorje Mts. in the SW and then continuing with the Graz Paleozoic, the basement of the Styrian Basin and the South-Burgenland swell in the E. In the northern part of the Eastern Alps, the Paleozoic of the Upper Austroalpine is represented by the Eastern Graywacke Zone which can be traced to the NE until the Vienna Basin (Fig. A.1).

The Gurktal nappe complex has two main nappe units composed of Lower Paleozoic rocks. The formations of the lower Murau nappe display an upper greenschist phase metamorphism while the overlying Stolzalpe nappe is metamorphosed in lower greenschist facies. The Ackerl crystalline with its gneiss and micaschist complexes forms a local klippe as the highest structural unit of the Upper Austroalpine in the Gurktal region.

The basal part of the Murau nappe comprises greenschists derived from mafic volcanites with slightly alkaline basalt characteristics (Neubauer, 1992). The overlying sequence is dominated by phyllites, carbonatephyllites and quartzites with smaller greenstone intercalations. Acidic volcanics are locally widespread (Loeschke, 1989). The upper part of the Murau nappe is characterized by carbonates, including marbles. Dolomites at the base of this carbonate succession were dated by Conodonts as Late Silurian to Early Devonian in age (Schönlau, 1979). The stratigraphy of the Murau nappe is comparable to that of the Schöckl nappe of the Graz Paleozoic (see later).

The Stolzalpe nappe is a stack of smaller imbricates displaying various Late Silurian to Early Devonian facies. The overall sequence, however, is comparable to that of the Murau nappe with some important differences (von Gosen, 1982; 1989). The mafic volcanites at the base can be found in two stratigraphic levels: the lower Magdalensberg Formation is late Middle Ordovician in age, while the upper Nock Series are of late
Ordovician age (Neubauer and Pistotnik, 1984). These Ordovician volcanics including the Kaserer Formation are separated from a Silurian sequence of volcanics by slates, quartzphylite, sandstones and lydites (Neubauer, 1980). The Silurian volcanics include the Eisenhut and the Metadiabase Formations. The carbonate succession in the upper part of the Stolzalpe nappe has dolomites and pelagic limestones locally as young as Carboniferous and is overlain by cherts and greywackes.

The Ackerl Crystalline is made up of a lower micaschist and an upper gneiss unit. The age of amphibolite facies metamorphism is pre-Alpine, but an Eoalpine overprint in lower greenschist facies can also be demonstrated (Dallmeyer et al., 1992).

The Graz thrust-complex is made up of three nappe systems. The lower is the Schöckl nappe which has three pre-Alpine units. The Schöckl and Passail successions consist of Upper Silurian to Middle Devonian rocks. The greenschist facies metamorphosed Upper Silurian to Lower Devonian sequence is characterized by metavolcanites and siliciclastics while the Middle Devonian developed in a carbonate facies. Amphibolite facies rocks can also be found in the Anger Crystalline complex in the eastern part of the Graz thrust-system (Neubauer, 1981).

The Laufnitzdorf and Kalkschiefer nappes of the middle nappe system have their own stratigraphy, namely the Laufnitzdorf and Kalkschiefer groups. The former consists of very low-grade to low-grade pelagic limestones, shales and volcanites spanning the Early Silurian to Late Devonian interval. The Kalkschiefer group is a homogeneous sequence of alternating carbonates and siliciclastics.

The upper two nappes have two successions, the Rannach and the Hochlantsch groups, respectively. Both have a Silurian to Carboniferous stratigraphic succession with a very low-grade to low-grade metamorphic overprint. In general, while mafic metavolcanites and clastics dominate the Silurian-Lower Devonian strata, the Upper Devonian-Carboniferous sequence has pelagic and platform carbonates.
In the basement of the Styrian Basin and the South-Burgenland swell, Kröll et al. (1988) distinguished a number of Paleozoic successions. The *Sausal group* is made up of epimetamorphic phyllites, calc- and sericite-phyllites and graphitic quartzites with carbonate and metabasalt intercalations. As part of the Stolzalpe nappe (see above) the Sausal group has a tectonic contact with the crystalline mass of Remschnigg. The age of this group is uncertain, ranging perhaps from the Ordovician to Silurian (Devonian?).

The *Radochen Schists* were encountered in the Radochen-1 borehole with 820 m apparent thickness. These are anchimetamorphosed, dark slates of uncertain age. Possible correlatives include the Eisenerz Schist of the Graywacke Zone, the Dult Slate of the Graz Paleozoic and the Hochwipfel Flysch of the Carnic Alps. Such correlations would indicate a Carboniferous age for these schists.

The *Wollsdorf Metabasite* formation is a sequence up to 370 m thick, containing metamorphosed mafics with accompanying tectonic breccias. It is correlated with the Silurian Metabasite of the lower Kher Schist (Flügel and Neubauer, 1984) and perhaps with the Paleozoic greenschists of Hannersdorf. Kröll et al. (1988) tentatively correlated this formation with the Sótony Metabasalt of the Hungarian part of the Danube Basin.

The *Blumau Phyllite-Carbonate* formation consists of phyllites, limestones and dolomites of Upper Silurian to Lower Devonian age. The outcrop of the formation near Sulz was dated as Upper Silurian by Schönlau (1984). The Blumau Phyllite-Carbonate is correlated with the Szentgotthárd Phyllite of Hungary, on the eastern flank of the South Burgenland swell (Balázs, 1983).

The *Arnwiesen Group* is made up of Lower Devonian grey dolomites at the base (Ebner, 1988). These rocks can be correlated with the Dolomite-sandstone succession of the Rannach nappe of the Graz Paleozoic (Ebner, 1978) and with the Bükk Dolomite formation of Hungary (Balázs, 1971). Other lithologies of the Arnwiesen group, such as sericite-quartzite, quartz- and sandstoneschist can be correlated with the Mihály Phyllite
formation. The dolomite is overlain by a 90 m thick black slate, which was correlated with the Arzberger formation of the Schöckl nappe of the Graz Paleozoic (Ebner, 1988).

The Eastern Graywacke Zone typically is made up of two nappes: the lower Veitsch nappe directly overlying the Middle Austroalpine and the upper Noric nappe. Locally at the base of the latter, slices of the crystalline basement are known as the Kaintaleck Complex. Recently, an additional nappe was distinguished between the Veitsch and Noric nappes, the Silbersberg nappe (Hermann et al., 1992).

The greenschist facies (Eoalpine) metamorphosed Veitsch nappe includes clastics and carbonates of Carboniferous age. Its internal stratigraphy was subdivided by Ratschbacher (1987) into three formations. At the bottom the Steilbachgraben Formation is made up of laminated siltstones with dolomite and magnesite lenses. The overlying Triebenstein Formation is characterized by massive carbonates. The uppermost Sunk Formation consists of a coarsening-upward sequence of alternating limestones, shales and conglomerates.

The new Silbersberg nappe has a thick base of chloritic calcschists of Early Paleozoic (?) age which is covered by a clastic sequence of metaconglomerates and metapelites. The clastic sequence has a continuous transition to an overlying phyllite succession. The thickness of the greenschist metamorphosed Silbersberg nappe is about 1500 m.

The Kaintaleck complex is a 140 m thick sequence of amphibolite phase rocks. This succession includes micaschists at the base overlain by amphibolites with marble intercalations. Although the whole unit has a pronounced greenschist facies metamorphic overprint due to Eoalpine overthrusting, dating of white micas yielded pre-Variscan ages.

The base of uppermost Noric nappe in the Graywacke Zone is dominated by the 400 m thick Blasseneck Porphyroid of Upper Ordovician age (Schönlaub, 1979). It is overlain by a thick sequence of schists and quartzites (Polster Quartzite). The Silurian part of the
sequence is characterized by a 200-400 m thick pile of mafic volcanites and limestones. The Devonian is dominated by a 200-300 m thick carbonate succession. Unconformably overlying them are thin layers of Carboniferous clastics that occur locally. The nonmetamorphosed Permo-Mesozoic succession of the Northern Calcareous Alps overlays the Graywacke Zone with the major Hercynian unconformity.

The Graywacke Zone can be traced along strike to the Vienna Basin (Fig. A.1) where the 800 m thick succession was drilled in numerous boreholes (Hamilton et al., 1990). The Graywacke Zone as the stratigraphic base of the Juvavic nappes seemingly pinches out to the NE just beyond the Austrian-Slovakian border, probably due to erosion.

The Choc, Lunz and the structurally higher other nappes in the Slovakian part of the Vienna Basin and in the northwestern Little Carpathians have an Upper Paleozoic base. The Ipoltica Group includes the Upper Carboniferous Nizná Boca Formation and the Permian Maluzind Formation. The latter is made up of a 700-900 m thick complex of basic volcanites. Coarse clastics and evaporites are intercalated within tuffitic parts of the section. The stratigraphic position of the formation is comparable to that of the Kékkút Dacite of the Bakony Mts.

A.2 EARLY-ALPINE (PERMIAN-LOWER CRETACEOUS) STRATIGRAPHY

This time interval is represented in all the units of the Austroalpine nappe system and the underlying Penninic units as well.

A.2.1 PENNICINC UNITS

The Rechnitz window group at the eastern end of the Eastern Alps comprises the Rechnitz, Berstein, Möltern and Eisenberg windows (Fig. A.1). The eastern parts of the Rechnitz and Eisenberg windows are in Hungary (see Chapter 4). The greenschist metamorphosed succession of the Rechnitz Series was first assigned to the Penninic unit
by Schmidt (1951).

In the probably overturned (?) sequence, metasediments are overlain by a tectonically reduced ophiolite series (Koller and Pahr, 1980). The more than 2000 m thick metasediments consist of calcareous micaschists and quartz phyllites. Their Mesozoic age was proved by fossils (Schönlaub, 1973). The lithology is equivalent to the classical Bündnerschiefer of South Penninic of the Central Alps. Additional lithologies are graphite phyllites, marbles, conglomerates and rauhwackes (Pahr, 1980). The ophiolite sequence, locally covered by some m thick radiolarites, consists of typically up to 200 m thick finely laminated greenschists. The plutonic sequence is represented by plagiogranites, ferrodiorites and ferrogabbros (Koller, 1985). Ultramafic rocks comprise completely serpentinized harzburgites, at least 260 m thick. Three metamorphic events were distinguished by Koller (1985). The first is an oceanic metamorphism in the Jurassic South Penninic ocean (Koller and Höck, 1990), followed by a Cretaceous low T/high P metamorphic event. This latter event and the final Miocene (22-19 Ma) greenschist facies metamorphic event were also recorded on the Hungarian part of the Rechnitz window group (see above).

The Borinka unit of the Little Carpathians has a particular lithology comparable to the typical South Penninic sequences of the Alps. It is also generally accepted that it was formed on the southern margin of a postulated oceanic basin. However, despite the ongoing debate, Plasienka et al. (1991) considered the Borinka unit to be still part of the Tatricum (Lower Austroalpine, see below).

A.2.2 LOWER AND MIDDLE AUSTRALPINE UNITS

The Permo-Triassic Semmering group starts with the Permian (?) Alpine Verrucano Formation (Tollmann, 1977). Lithologically it comprises sericite schists, breccias, tuffs and minor andesitic extrusive rocks. The overlying Semmering Quartzite and grey slates
are regarded as Late Permian to Scythian in age. The Middle Triassic carbonates consist of rauhwackes at the base which grade into limestones and dolomites and finally into dark slates. The Upper Triassic Carpathian Keuper (Tollmann, 1977) with its terrestrial, variegated clays is markedly different from the Central Alpine Triassic facies. The cover of the Wechsel/Waldbach units of the Lower Austroalpine is similar to that of the Semmering but it ranges only up to the Anisian, perhaps due to tectonic omission.

The low grade Stangalm Permo-Mesozoic transgressively overlies the Middle Austroalpine crystalline at the northwestern margin of the Upper Austroalpine Gurktal Complex. The Stangalm Mesozoic has a Central Alpine character. Permian sericite-quartzites and quartzkeratophyres are overlain by Scythian quartzites, Anisian rauhwackes, well-bedded limestones, Ladinian Wetterstein Dolomite and Carnian slates on top.

The early Alpine sediments of the Tatricum in the Little Carpathians are grouped into several formations (Plasienka and Putis, 1987). While the Jurassic and Lower Cretaceous strata can be subdivided into numerous formations based on their large lithological variabilities, the Triassic sequence shows a quite uniform development of continental and shallow marine facies.

Sedimentation began on the eroded Hercynian basement with terrestrial clastic deposition of the Devín Formation during the Late Permian. The overlying alluvial sandstones of the Lower Triassic Lúzna Formation form a widespread, 100-200 m thick sedimentary blanket. Variegated "Werfen" shales follow. The overlying 500 m thick Middle Triassic is made up of platform carbonates, limestones and dolomites. Upper Triassic strata are missing due to erosion, but they probably were represented by continental beds of the Carpathian Keuper.

Pronounced extensional tectonics during the Early Jurassic led to the development of distinct lithofacies units deposited in deep troughs that were separated by shallow
submarine highs. Six units were distinguished by Plasienka et al. (1991): *Borinka, Oresany, Devín, Kuchyna, Kadlubek* and *Solírov successions*. The Borinka succession developed on the SE flank of an oceanic domain (South Penninic or Vahic of Slovakian authors). This ocean was separated by a continental area (analogous to the Lungau Swell of the Eastern Alps) from a southerly trough (Siprun/Fatra zone), where other Jurassic successions developed.

The *Borinka Formation* is made up of scarp breccias derived from active fault zones. The Lower Jurassic *Prepadlé Formation* is characterized by the dominance of Triassic carbonate clasts (up to 300 m thick), while the 500 m thick Middle to Upper Jurassic *Somár Formation* is dominated by crystalline clasts. These breccias are comparable to those which were described from the Rädstatter Tauern of the Eastern Alps (Häusler, 1987) and also from the Lower Austroalpine unit. Farther to the NW, in a more basinal setting during the Lower Jurassic, the turbiditic *Korenec Formation* was deposited (up to 800 m). In the axis of the active half-grabens the 500 m thick hemipelagic *Marianka Shale* of Toarcian-Bathonian age formed (Plasienka, 1987).

To the S of the continental ridge markedly different Jurassic lithofacies developed. Deposition of calciturbidites (500-600 m thick) in the Oresany unit indicates the formation of a relatively deep trough during the Toarcian(?) which was separated from other successions in the Siprun Zone. The *Devín Formation* group is characterized by neritic Lower Jurassic and bathyal Middle Jurassic to Lower Cretaceous carbonates. The Kuchyna succession has a similar character, but the Lower Jurassic overlies directly the crystalline basement. The Solírov succession has a specific Barremian re-deposited bioclastic sandy limestone unit (*Solírov Formation*). The *Kadlubek Formation* group with its reduced thickness (50 m) displays the characteristics of an intra-basinal high with condensed sedimentation.

This Jurassic paleogeographic pattern played an important role during the
subsequent Eoalpine movements. The Jurassic listric faults bounding the continental swell were reactivated as overthrust planes forming the Modra and Bratislava basement nappes (Plasienka et al., 1991).

The early Alpine strata of the Vysoká Nappe of the Fatricum was deposited just S of the Tatric units described above. The Anisian Vysoká Formation is about 200-250 m thick and has a Gutenstein-type, carbonate ramp facies. The overlying Ramsau Dolomite is 40-60 m thick. Its upper part is characterized by limestones of Carnian age comparable to the Opponitz beds of structurally higher units. The Carpathian Keuper with its red and variegated shales is up to 300 m thick in several places and has intercalations of grey dolomites in its upper part. The Upper Rhaetian Fatra Formation is a sequence of neritic fossil-rich limestones, overlain by the 100 m thick, shaly Kopieniec Formation of Hettangian age. The Middle Jurassic strata include dark shales and cherty limestones. The Upper Jurassic sequence is represented by the Niedzica and Czorsztyn Formations in Ammonitico Rosso facies. The intercalated Czajakowa Formation radiolarites are Oxfordian in age.

The pelagic Lower Cretaceous succession is represented by the Padlá Voda Limestone and the marly limestones of the Hlboc Formation. The deep-water strata are covered by the neritic, organogenic limestones of the Bohát Formation.

A.2.3 UPPER AUSTROALPINE UNITS

In the region of the Gurktal nappe complex (Fig. A.1), early Alpine cover sequences are found in two areas. The Pfannock sequence at the NW part of the Gurktal complex is a detached portion of the Stolzalpe nappe and has a Triassic facies similar to that of the Middle Austroalpine Stangalm unit (see above). In the Krappfeld area, the Eberstein Permotriassic succession consists of a basal terrestrial unit (Werczheim Formation) which grades upward into the sandstones of the Gröden Formation. The Anisian rauhwackes are
overlain by an incomplete and tectonically disturbed Upper Triassic sequence with clastic Carnian Raibl Beds and the Norian Hauptdolomit.

In the pre-Tertiary basement of the Styrian Basin at the South-Burgenland swell, the Radkersburg borehole (Fig. A.1) penetrated a Permotriassic sequence, perhaps in an Upper Austroalpine position. The Radkersburg group of Kröll et al. (1988) comprises a 30 m thick Permian (?) and a 67 m thick Middle Triassic (?) carbonate sequence. In the nearby Remschnigg area of the Pohorje Mts., a comparable Permotriassic sequence is known from an imbricate slice. This includes the 300 m thick Permian (?) sandstone unit at the base followed by a 100 m thick brecciated Middle Triassic (?) dolomite and 500 m thick Upper Triassic (?) grey dolomites on top (Mioc., 1977).

The eastern end of the Northern Calcareous Alps and its continuation in the pre-Tertiary basement of the Vienna Basin is subdivided into three Upper Austroalpine nappe systems (e.g. Wessely, 1988). The lowermost Bajuvaricum is represented by the Frankenfels-Lunz unit and the Tirolicum by the Göller nappe, while the uppermost Upper Limestone Alpine nappes belong to the Juvavicum.

In the Triassic succession of the Frankenfels-Lunz unit the sequence starts with the Upper Anisian to Upper Ladinian Reifling Limestone which can attain a thickness up to 480 m (Hamilton et al., 1990). The Lower Carnian Lunz Beds are characterized by 600 m thick coal seams. The overlying 80 m thick Carnian Opponitz Beds are made up of dolomites and evaporites. The Norian Hauptdolomit consists of 500-1000 m thick grey dolomite with shale interbeddings. The mass of the Hauptdolomit is overlain by the Rhaetian Kössen Beds, a lagoonal, organic-rich shaly sequence of 50 m thickness. The 20 m thick Liassic Hierlatz Limestone indicates deposition in shallow water. The overlying Dogger Klaus Limestone and the Malm radiolarites and pelagic limestones are not more than a hundred meters thick. In the Neocomian 200 m thick marls and marly limestones occur (Schrambach Beds).
The early Alpine sequence of the Göller nappe starts with the Permian evaporitic Haselgebirge and the Permocyclyan clastic Werfen Beds. The overlying Anisian Reichenhall and Steinalm Beds and the Gutenstein Limestone is 300 m thick. The upper Middle Triassic lithofacies of the Göller nappe is subdivided into two types based on the presence of basinal or platform limestones (Reifling and Wetterstein Limestone, respectively). The 150 m thick Reifling Beds have marls, shales, cherts and tuffs. These lithofacies change to the SE into several hundred m thick Wetterstein Limestone. The Carnian Lunz Beds are overlain by the up to 1700 m thick Hauptdolomit. In the southern part of the Göller nappe, the Kössen Beds and the Hauptdolomit are replaced by the Norian 100 m thick Plattenkalk and the 350 m thick Dachstein Limestone. In the same area the 500 m thick Jurassic is made up of the Liassic spotted limestone of the Adnet and Allgäu Beds followed by radiolarites and Oberalm Beds.

Only the Triassic succession is known from the higher Limestone Alpine nappes and it is similar to that of the Göller nappe. The Middle and Upper Triassic platform carbonates (Wetterstein and Dachstein Limestones) both attain a thickness of 1300 m in this nappe system (Hamilton et al., 1990).

In the Slovakian part of the Vienna Basin and in the Little Carpathians, the early Alpine succession is similar to that of the Austrian side of the Vienna Basin. The Lower Triassic Benkovsky Potok and Sunava Formations lie transgressively on a Hercynian (Carboniferous) basement. The 200-250 m thick shallow marine siliciclastic sequence displays an overall upward fining character. Deposition of carbonates began in the Anisian (early Pelsoian). The basal member of the Gutenstein Formation consists of bituminous, oolitic micrites deposited in a very shallow, hypersaline environment. The overlying Annaberg Limestone is about 200-300 m thick. The limestones of the Steinalm Formation have a variable thickness; they are 400-500 m thick in the Havranica and Jablonica nappes, 30-40 m in the Veterlin nappe and only a few m in the Choc nappe, indicating the growth
of an extensive carbonate platform. The Upper Anisian Zámostie Formation of the Choc and Veteranin nappes represents carbonates deposited in intraplatform depressions.

The Ladinian Reifling Formation comprises a 100 m thick cherty, basinal limestone sequence. In the Havranica and Jablonica nappes the heterotrophic succession is the 800 m thick Wetterstein Formation. The Upper Ladinian Raming Limestone is an allogapic limestone unit in the Veteranin nappe. The Upper Triassic sequence comprises the Reingraben Formation with intercalated Lunz clastics and the overlying Opponitz Limestone capped by thick Hauptdolomite. Depending on the amount of Late Cretaceous erosion locally the uppermost part of the Triassic is still preserved; in the Havranica nappe the Dachstein Limestone and the Rhaetian Norovica Formation form the top of the Early Alpine succession. Jurassic or Lower Cretaceous formations are not known in this area, because of subsequent erosion.

A.3 EOALPINE STRATIGRAPHY (MIDDLE CRETACEOUS - PALEOCENE)

This time interval is represented only in the Austroalpine nappe system. The lower boundary of this period is drawn at the beginning of the Albian. The upper boundary is poorly defined and is regarded as the Paleocene/Eocene boundary.

The Styrian Gosau basins (Kainach, Krappfeld and St. Paul) are located in the Central Alpine zone (Fig. A.1). The Middle Styrian Kainach Gosau overlies the SW part of the Graz Paleozoic. In the northern part of this Senonian basin, the 600 m thick basal conglomerate (Basisconglomerate) was deposited as an alluvial/fluvial succession. The composition of these clastics is of special importance since it contains various exotic Mesozoic rocks of South Alpine origin (Gollner et al., 1987). The conglomerate is overlain by a 200 m thick limnic sequence (Bitumenmergel) encountered in the Afling U 1 well (Kröll and Heller, 1978). An Early Campanian 1200 m thick flyschoid sequence follows (Hauptbecken-Folge). In the SE part of the Kainach area, the topmost part of the
Senonian is a 250 m thick Maastrichtian marl and calcarenite (*Zementmergel*).

The Krappfeld Gosau basin contains a 1500 m thick Senonian sequence (Van Hinte, 1963; Neumann, 1989). The basal transgressive beds overlie the Hauptdolomite and part of the Paleozoic Gurktal Complex. Detrital reef limestone with *Hippurites* can be found locally above the basal beds. The overlying pelagic sequence has a flysch character with large (up to 30 m thick) olistolith intercalations (Neumann, 1989). The St. Paul Basin shows a similar sequence but it is erosionally truncated on top.

The Gosau Basins of the eastern part of the Northern Calcareous Alps (Faupl et al., 1987) are located in isolated patches as erosional remnants of an originally continuous basin. Some of these small basin fragments are shown in Fig. A.1. The Gosau sequence is subdivided into a lower and an upper complex. The Coniacian to Lower Campanian lower complex comprises a basal conglomerate of alluvial origin called *Kreuzgraben Formation*. This is overlain by the transitional terrestrial to shallow water *Streitck Formation*. Neritic and deeper water fossiliferous marls (*Schneckengraben* and *Grabenbach Formations*, resp.) are interfingering with the rudistic limestones of the *Hochmoos Formation*.

All of these shallow marine formations are separated from the overlying bathyal upper Gosau complex (*“Flyschgosau”*) by a pronounced erosional unconformity. This unconformity is either overlain by bathyal marls with distal turbidite intercalations (*Nierental Formation*) or by a thick (up to 240 m) carbonate breccia unit (*Spitzenbach Formation*). Locally, the Nierental Formation is underlain by a deep water sandy unit called the *Ressen Formation*. The Maastrichtian to Paleocene formations of the upper part of the "flyschGosau" were deposited below the CCD (Faupl et al., 1987). In the Weyerer Bögen, the thickness of the atypical turbidites of the *Brunnbach Formation* reach 1000 m. The turbidites of the *Höllgraben Formation*, however, show all the characteristics of the Bouma sequence. The deep water channel-filling clastics of the *Wörschachberg* and *Zwiesealm Formations* represent more proximal facies than the previous formations.
In the Frankenfels-Lunz nappe of the Vienna Basin overlying the Neocomian unconformity, the Upper Aptian to Albian *Losenstein Beds* consists of 250 m thick marls and sandstones. The 100 m thick Cenomanian *Sparbach Formation* unconformably overlies the underlying Cretaceous sequence with basinal marls and with clastics at the basin margins. Finally, the 80 m thick Turonian is represented by brackish, limnic and fluvial clastics with coal beds.

The Senonian to Paleocene *Gosau* succession of the Gießhübl syncline consists of Coniacian to Santonian basal breccias and sandstones. The overlying Campanian carbonates and marls show a gradual deepening of the basin. Upper Maastrichtian to Middle Paleocene marls with turbidites attain a thickness of 1400 m. The Upper Cretaceous succession of the Glinzendorf syncline shows a different facies. The *Gosau* succession in this area developed in a nearly 1000 m thick limnic facies, unconformably overlying Triassic rocks.

In the Slovakian part of the Vienna Basin, the Myjava syncline contains a Paleocene flysch succession several hundred m thick. The similar lithologies suggest that the Myjava syncline may have been connected with the Gießhübl syncline of the Austrian part of the basin (Hamilton et al., 1990).

In the Little Carpathians the Senonian post-tectonic cover is comparable to the *Gosau* sequence of the Eastern Alps and begins with the basal conglomerates of the Coniacian *Valchov Conglomerate*. The overlying 50-150 m thick *Baranec Sandstone* and the pelagic Upper Coniacian *Stvernik Formation* with its pelagic marls show the rapid deepening of the Senonian sea to bathyal water depths. The overlying Santonian *Hurbanova Formation* is an extremely thick (3000 m) flysch succession. The Lower Campanian *Kosaríská Formation* includes variegated marls. The Upper Campanian to Maastrichtian *Podbradlo and Bradlo Formations* have a flysch lithology with high carbonate content and sandstone intercalations. In the NW part of the Little Carpathians,
the Bartalová and Kriz'la Formations comprise red breccias which represent the heterotrophic, continental facies of the above described Senonian succession.

A.4 MESOALPINE STRATIGRAPHY (EOCENE - LOWER MIocene)

Since the rocks deposited during this time interval postdate the main formation of the Alpine nappe system, the allocation to Austroalpine units would be meaningless. Thus the description below follows geographic units.

In the area of the Gurktal Complex at Krappfeld (Fig. A.1), an uppermost Paleocene to Middle Eocene succession is known (van Hinte, 1963; Wilkens, 1989). The present-day distribution of these Paleogene rocks and those of the underlying Gosau sequence is largely controlled by a Neogene graben system. Thus the Krappfeld Paleogene represents the erosional remnant of a much larger basin. The 10 m thick Paleocene basal beds comprise red clays, gravels and local coal seams above an erosional unconformity. The overlying marls and limestones with abundant *Foraminifera* show the gradual deepening of the basin. The continuous 200 m thick Eocene succession ends with Lutetian limestones which are erosionally truncated (Wilkens, 1989).

In the northern part of the Little Carpathians in the Buková trough (Fig. A.1) an erosional remnant of an extensive Paleogene succession is preserved (Gross and Köhler, 1989). The conglomerates, breccias and bioclastic limestones of the 80-100 m thick *Borové Formation* show a Late Paleocene transgression on the eroded, locally karstified Triassic of the Upper Austroalpine nappes. The Lower to Middle Eocene *Huty Formation* was deposited in slightly deeper water, in neritic environments. The 400-500 m thick formation consists mainly of shales with sandstone and conglomerate intercalations. The enigmatic *Hrabnik Formation* consists of dark clays and siltstones with turbiditic sandstone beds. The lithology of this formation is comparable to that of the Zdanice unit of the Flysch Zone. After a hiatus the 20-50 m thick Lower Miocene (Eggenburgian)
Dobrá Voda Conglomerate records the transgression of the sea in the Little Carpathians area (Kovác et al., 1989). The overlying Luzica Formation with its dark shales indicate euxinic conditions.

In the Slovakian part of the Danube Basin, in its southeastern part Eocene to Oligocene rocks are known from wells and also from outcrops (Samuel, 1973). These rocks are essentially identical to those of the Hungarian territory described in Chapter 4.

A.5 NEOALPINE STRATIGRAPHY (MIDDLE MIOCENE - RECENT)

Very thick sedimentary successions were formed in several basins during the last stage of the Alpine evolution at the junction of the Eastern Alps, the Western Carpathians and the Pannonian Basin.

In the Styrian Basin (Fig. A.1), Middle Miocene sedimentation started in the Karpatian (Kollmann, 1965). The 200 m thick Styrian Siltstone ("Schlier") was deposited in bathyal (>100 m) water depth (Friebe, 1990) in the Gnas subbasin. At the same time to the NE, in the Fürstenfeld subbasin conglomerates formed a fan delta. In the western part of the Styrian Basin, the fluval and paralic(?) Eibiswald Formation was deposited with thick coal seams during the Karpatian. The continental to shallow-water western part of the basin was connected with the deeper water eastern part by the deltaic Arnfels Conglomerate. The Karpatian sequence is terminated by the important Styrian angular unconformity (Friebe, 1991). In the Styrian Basin, several Middle Miocene (Karpatian-Badenian) volcanic centers are known. The mainly trachyandesitic volcanism with its lavas and tuffs shows a temporal change from subalkaline to alkaline character (Ebner and Sachsenhofer, 1991).

The Badenian consists of neritic coralline algal limestone of the Leithakalk and siliciclastics (Weißenegg Formation). During the end of the Badenian in the western Styrian Basin, the braided delta of the Eckwirt Gravel prograded to the E. The Sarmatian
*Waldhof Formation* is comprised of interfingering shallow-marine and fluvial sediments near Graz (Fig. A.1). Similarly siliciclastics and carbonates are interfingering (*Rollsdorf Formation*) in the northern part of the Styrian Basin.

The Upper Miocene (Pannonian) *Weiz Formation* with its coal seams was deposited as a prograding braided delta in the northern part of the eastern Styrian Basin. The overlying *Puch Gravel* and *Kapfstein/Kirchberg Gravel* formations correspond to alluvial fan and meandering river sedimentation, respectively, as the Styrian Basin was overfilled. The Pliocene-Pleistocene volcanic activity resulted in approximately 40 eruptive centers in the central part of the Styrian Basin. The lava flows and pyroclastic rocks are of alkaline composition (nephelinitic/basanitic, Ebner and Sachsenhofer, 1991).

The Vienna Basin has a very thick (up to 5 km) Neogene sedimentary fill. In the northern part of the basin where it was connected with the Molasse basin (Fig. A.1), there is no break in sedimentation between the Early and Middle Miocene (e.g. Jiricek and Seifert, 1990). The southern part shows a comparable development to that of the Pannonian Basin; however, sedimentation started in the Middle Miocene. The 300 m thick basal conglomerates, terrestrial sandstones and shales of the lower Karpatian *Gänsendorf Beds* are overlain by the 800-1000 m thick upper Karpatian *Aderklaa Beds*. This latter succession comprises limnic to fluvial conglomerates, sandstones and marls. There is a pronounced, locally erosional unconformity at the Karpatian/Badenian boundary. The overlying marine Badenian sequence is up to thousands of meters (up to 3500 m) thick depending on its tectonic position relative to major normal faults (Wessely, 1988). The top of the sandy and locally carbonatic Badenian shows brackish influence, which becomes even more pronounced in the Sarmatian sequence. The up to 1000 m thick Sarmatian has less dramatic thickness changes than the underlying Badenian. The max. 1400 m thick Upper Miocene to Pliocene (Pannonian) is characterized by alternating lacustrine marls and sandstones in its lower part, while higher in the section sandstones prevail. Locally, in
the still subsiding parts of the basin the thickness of Quaternary fluviatile clastics reaches 200 m.

Sedimentation was continuous in the Little Carpathians at the Early and Middle Miocene boundary (Otnangian/Karpathian). The Lower Karpatian Laksár Formation is characterized by a bathyal siltstone ("Schlier") with frequent turbidite intercalations. The polymict, 300-500 m thick Upper Karpatian Jablonica Conglomerate was deposited in an alluvial delta fan (Kovác, 1985). On the western side of the Little Carpathians, at the eastern margin of the Vienna Basin, the Middle Miocene (Middle Badenian) Devínska Nová Ves Formation points to the opening of the basin. The 300-350 m thick formation is made up of conglomerates and breccias with large boulders indicating sedimentation in a talus adjacent to an active fault zone. The Upper Badenian Sandberg and Studienka Formations comprise sandstones and marls, respectively, depending on their relative position to the basin margin.

In the northern part of the Danube Basin, sedimentation started in the Middle Badenian. The transgressive Dol'any Conglomerate has a polymict character containing pebble material of local sources. The sands and clays of the Madunice Formation contain abundant macro and microfauna. The overlying Neogene succession is known only from boreholes in the basin area and is comparable to that of the Hungarian part of the Danube Basin.

A.6 DÉCOLLEMENT LEVELS IN THE PALEozoIC AND MesoZOIC OF THE BAKONY MTs., DANUBE AND ZALA BASINS.

In the Paleozoic succession, several potential décollement horizons can be delineated based mainly on borehole data. The Upper Ordovician Balatonfökajár Quartzphyllite was found above the Devonian Polgárdi Marble with a thrust contact in the Szabadbattyán-9 well (Fig. A.2) at 92 m depth (Fülöp, 1990). In the same well, the Devonian marble
Fig. A.2. Index map of the study area in the NW Pannonian Basin.
tectonically overlies the Lower Carboniferous Szabadbattyán Schist between 240-320 m. In the Úrhida-4 borehole (Fig. A.2), the Silurian Lovas Schist has a brecciated tectonic contact at 50 m depth with the underlying Devonian Úrhida Limestone (Füllöp, 1990).

In the Mesozoic, the Lower Triassic Arács Marl with its evaporite intercalations may be an excellent detachment surface (Fig. A.3). In the Dióskál-5 borehole just to the W of the Lake Balaton (Fig. A.2), in a Permoscythian succession several repetitions were observed (Körössy, 1988). Between 931-1201 m red sandstones, clays (Balatonfelvidék Sandstone) were drilled followed by anhydrite, oolitic dolomite and anhydritic dolomite (Arács Marl) between 1201-1305 m. Between 1305 and 1385 m again red sandstones were penetrated followed by anhydritic layers between 1385-1410 m. The underlying red sandstone was drilled between 1410-1541 m and to the bottom of the borehole again anhydritic sandstones, claymarls and dolomites were found. These repetitions were explained in terms of small-scale imbrications by some; however, Körössy (1988) proposed an alternative view in which the red continental beds are simply intercalated with lagoonal evaporites. The contact between the Scythian Hidegdút Sandstone Member of the Arács Marl and the overlying Csopak Marl was also found to be sheared in the Bakonyszűcs-3 (Bence et al., 1990) and in the Györszemere-2 (Teleki et al., 1989) boreholes (Fig. A.2).

The base of the Anisian Megyehegy Dolomite seems to be an important detachment surface (Fig. A.3). Teleki et al. (1989) described a brecciated base of this dolomite in the Györszemere-2 well where it has an apparently tectonic contact with the underlying Aszőfő Dolomite (the Iszkahegy Limestone is missing!). The tectonic omission is even more pronounced in the nearby Györszemere-3 well (Fig. A.2). The Megyehegy Dolomite was found on top of the Paleozoic Tét Schist. Furthermore, in the Zebecke-3 well in the Zala Basin (Fig. A.2), an Anisian(?) dolomite was found between 2652-3478 m which tectonically overlies strongly sheared Carnian marls down to the bottom of the well at
Fig. A.3. Potential décollement levels in the Triassic of the Bakony Mts.
3978 m (Körössy, 1988). Dudko (1991) observed a layer-parallel thrust contact between the Megyehegy Dolomite and the Iszkahegy Limestone in the vicinity of Balatonfüred (Fig. A.2). This contact was termed the Southern thrust (see on Fig. 4.14) and it may indicate a thrust flat.

The Ladinian Buchenstein beds with their incompetent tuff and clay layers are also reported to be associated with thrusting (Fig. A.3). In the Iszkahegy area (Fig. A.2), the Bakonykuti thrust follows this level as it was revealed by surface mapping (Raincsák, 1980).

One of the most important décollement levels can be found within the Carnian Veszprém Marl and especially at its contact with the overlying Norian Hauptadolomit (Fig. 6.2). The best known example is that of the Veszprém thrust (Laczkó, 1911; Peregi and Raincsák, 1983; Dudko, 1991). Near the town of Veszprém (Fig. A.2), a Lower to Upper Triassic sequence is overthrust on top of the Hauptadolomit of the footwall with a westward decreasing stratigraphic gap. Similarly, the anomalously thick Hauptadolomit strip to the N of Veszprém can be explained only by several internal thrusts (see Chapter 5, cf. also Haas and Jocha-Edelényi, 1980; Peregi and Raincsák, 1983; Dudko, 1991). Another field example is given by Gyalog and Raincsák (1981), who found a thrust contact between the Dachstein Limestone and the Hauptadolomit in the area of Bakonyszentlászló (Fig. A.2). The thrust plane (110°/28°) here represents a thrust ramp since the underlying Dachstein Limestone shows an almost subhorizontal dip (2°). Bencze et al. (1990) also mentioned the tectonic contact between the Veszprém Marl and the Hauptadolomit in the northern Bakony.

In the subsurface of the Zala Basin, the Bárszentmihályfa-I well (Fig. A.2) penetrated the Hauptadolomit between 2982-4451 m in a tectonic contact with a Lower Jurassic (Sinemurian?) Hierlatz-type limestone (Körössy, 1988). In the Györszemere-2 well, the Norian Hauptadolomit has a brecciated contact with the underlying Anisian
Megyehegy Dolomite interpreted by Teleki et al. (1989) as a thrust contact. In some wells of the area the Hauptdolomit turned out to be anomalously thick indicating internal repetitions. The Nagylengyel-108 well (Fig. A.2) penetrated the Hauptdolomit between 2335-4409 m which is usually only 800-1200 m thick. Similarly, stratigraphic repetition by thrusting can be supposed in the case of the Bakonyszücs-1 well (Fig. A.2), where a 1153 m thick sequence of the Veszprém Marl was drilled (Bence et al., 1990).

The marls of the Kössen Formation or other transitional layers between the competent layers of the Hauptdolomit and the Dachstein Limestone can be regarded as the uppermost decollement level for the Eoalpine compressional period (Fig. A.3). In the area of Sümeg (Fig. A.2), Knauer and Gellai (1978) considered the contact between the Kössen Formation and the Hauptdolomit to be tectonic. In the Bárszentmihályfa-I well, three imbricates were drilled below thick Hauptdolomit strata (Körössy, 1988). The uppermost imbricate between 4451-4550 m shows a normal stratigraphic sequence ranging from a Sinemurian(?) limestone to the Rhaetian Dachstein Limestone. The intermediate imbricate between 4550-4850 m is made up of Pliensbachian black marls, Sinemurian-Hettangian nodular limestones to Rhaetian Dachstein Limestone. Finally, from 4850 m to the bottom at 5075 m Bathonian-Callovian black limestones, Aalenian-Toarcian Ammonitico Rosso and finally Pliensbachian-Sinemurian limestones were penetrated. In the Zebecke-2 well (Fig. A.2), between 2532-2814 m Rhaetian Dachstein Limestone and Kössen Formation were found on top of Middle Eocene(!) sediments between 2814-3348 m (Körössy, 1988). The Eocene strata exclude the Eoalpine age of thrusting, but from a mechanical point of view the well still suggests that the Kössen beds define a potential decollement level (Fig. A.3).
APPENDIX B
TERTIARY SEQUENCE STRATIGRAPHY OF THE PANNONIAN BASIN

B.1. THE POST-RIFT SUCCESSION (UPPER MIOCENE - RECENT).

Seismic stratigraphic (Mattick et al., 1988) and sedimentological studies (Bérczi, 1988) in southeastern Hungary indicate that the post-rift infill of the Pannonian Basin involved the advance of deltas from the northwestern, northern and northeastern basin margins. Deltas developed in two stages. The first stage deltas prograded into a 800-900 m deep lake; during this phase the rates of subsidence and sediment supply were very high. Sedimentation locally exceeded rates of 1000 m/Ma. Following a "destructive phase" (Mattick et al., 1988), delta construction during the second stage was characterized by progradation in shallower water (200-400 m). Finally the rate of sediment supply exceeded the rate of thermal subsidence, resulting eventually in complete infilling of the Pannonian lake.

Seismic line Qb given in Fig. 3.34a, is located in eastern Hungary (Fig. 3.1). In this particular region the prograding delta-succession corresponds to the older-stage of delta construction (cf. Mattick et al., 1988, their Fig. 20). The analysis of characteristic reflection termination patterns such as downlap, onlap, toplap, sigmoidal and oblique offlap allows a sequence stratigraphic interpretation of the profile, according to seismic stratigraphic mapping techniques described by Vail (1987) (Fig. 3.34b).

Based on stratal patterns, several depositional sequences (Posamentier et al., 1988 and Vail et al., 1991) can be recognized in the section (Fig. 3.34b). All of them are characterized by erosion on the delta plain and pronounced downlap and toplap at the base and at the top of the highstand systems tracts, respectively. The lowstand systems tracts are not fully developed except for the sequence on the southwestern part of the section. The sand sheet of the basin floor fan and the overlying slope fan mounds indicate a major
fall in the water level of the Pannonian lake. Mattick et al. (1988) mapped this sequence and interpreted it as the result of a "destructive phase" in the delta construction, most probably caused by a shoaling of the lake.

Pogácsás et al. (1988) also recognized some pronounced unconformities in the post-rift sedimentary succession in northeastern Hungary and postulated that although the Pannonian lake was totally isolated from the world oceans from 10.5 Ma onward (Steininger et al. 1988), these unconformities show a strong correlation with the eustatic curve of Haq et al. (1987). Since the standard biostratigraphy cannot be used in lake sediments, Pogácsás et al. (1988) calibrated their seismic lines with magnetostratigraphy recorded in key wells and thus correlate four unconformities in the Pannonian Basin with the four third-order late Miocene eustatic sea-level minima of the global curve of Haq et al. (1987).

Taking into account that during this time interval the progradation of the Pannonian delta system was extremely rapid, about 30 km/Ma, the average duration of each sequence in Fig.3.34b is estimated as 0.1-0.5 Ma. Therefore they exemplify fourth-order sequences which are simple transgressive-regressive sequences (Vail et al., 1992). Fourth-order sequences tend to occur in sets which form what is defined as a third-order composite sequence (Van Wagoner et al., 1989). The fourth-order sequences are believed to be caused by climatic fluctuations associated with Milankovitch scale orbital cycles, namely with the 400 ka and 100 ka eccentricity cycles (Fischer, 1986). Therefore, the small amplitude lake-level variations (on the order of perhaps some meters) of the Pannonian lake, indicated by the sequence stratigraphical interpretation of seismic section Qb, were caused most probably by climatic changes in the drainage areas of the rivers (i.e. in the uplifting Carpathians) flowing into the Pannonian lake. Fourth-order sequences can be identified seismically only under very favorable conditions, as pointed out by Fulthorpe (1991). Decreasing subsidence and also an extremely high rate of sediment supply during
the post-rift phase of the Pannonian basin resulted in an exceptional sequence resolution ("mature phase" of Fulthorpe, 1991).

However, since the depositional sequence associated with basin floor and slope fans (Fig.3.34b) indicates a major fall in lake level that can be mapped regionally (Mattick et al., 1988), I propose that this particular event represents a third-order sequence boundary (Tari et al., 1992a,b). Since Pogácsás et al. (1988) established a correlation between similar unconformities in the nonmarine sedimentary fill of the Pannonian Basin and the eustatic curve of Haq et al. (1987), the question arises: what is the reason for this coincidence? I will discuss two alternative explanations below.

The proposed correspondence between Vail's third-order cycles and the unconformities in the upper Miocene non-marine depositional sequences of the Pannonian Basin (Pogácsás et al., 1988) offers an excellent opportunity to address the problem of the water-level fluctuations of large inland lakes. Posamentier and Vail (1988) hypothesized that eustatic changes affect all streams that drain to the sea, producing non-marine sequences that correlate with marine counterparts. Therefore one possible explanation for the strong correlation is that the water-level variations of the Pannonian lake were influenced by eustasy, since the global sea level was the ultimate base level to which a large river connecting the lake to the sea adjusted itself. A "proto-Danube" of this type indeed could exist during this time interval, since the sudden decrease of salinity at the base of the post-rift succession (10.5 Ma) indicates high fresh-water input and a positive water balance in the Pannonian Basin (see Kázmér, 1990). The inflow which resulted from rivers draining the surrounding Carpathians was greater than evaporation; therefore an outflow existed during the late Neogene probably to the Aegean (Tari et al., 1992a).

In one particular case this mechanism appears to be effective. The major third-order boundary shown in Fig. 3.34 was dated as about 6.3 Ma (G. Vakarcs, pers. comm.). Therefore I speculate that the corresponding large (200-400 m) fall in the lake-level is the
"far-field effect" of the Messinian crisis in the Mediterranean (Hsü et al., 1977, 1978; Cita et al., 1978; Cita and Corselli, cum. lit.). My tentative model, which also takes into consideration the dessication of the Black Sea (Hsü and Giovanolli, 1980), is shown in Fig. B.1. During the pre-Messinian times the Pannonian Lake drained into the Black "Sea", since a divide provided by the Dinarides separated it from the Mediterranean Sea (Fig. B.1a). The Black Sea was actually a large inland lake with a steady water balance. Since the Mediterranean dessication caused the rapid headward erosion of rivers draining into the area of the present-day Adria (Lago Mare), the Pannonian Lake was reached and drained by a major river (Fig. B.1b). Turned off from the water supply of the Pannonian area the Black Sea was characterized by a negative water balance and thus a dessication event occurred. This salinity crisis, however, was not as dramatic as in the Mediterranean region (Hsü and Giovanolli, 1980). I believe that the most likely course for the "proto-Danube" is in the Vardar zone where the Pannonian water mass drained into the Aegean. Note that this possibility was already mentioned by Hsü et al. (1978).

Alternatively, one should take into consideration the effect of temporal fluctuations in intraplate stress levels, which can also create short-term variations in erosional base level (Cloetingh et al., 1985), similar to those which are attributed to glacio-eustatic processes (Bartek et al., 1991). Recent work by Cloetingh et al. (1989) indicates that Vail's coastal onlap curves are strongly influenced by tectonic events. Indeed, during the given time period large-scale rotations of the paleo-stress field occurred in the Mediterranean region (Philip, 1987) and also in the Pannonian basin (Csontos et al., 1991), suggesting that the state of stress could vary on a relatively short time scale, comparable to Vail's third-order cycles (1-5 m.y.). Additional evidence for intraplate stress variation in the region can be inferred from irregularities of the subsidence history during the post-rift phase and the study of the recent stress field (Dövényi and Horváth, 1990, Gerner, 1992).
Fig. B.1. The impact of the Messinian event on the Pannonian Lake.
Another commonly overlooked factor in the generation of depositional sequences is the rate of sediment supply. Since the source areas of sediments deposited in the prograding deltas were located in the surrounding and rapidly uplifting Carpathians, this indirect tectonic effect must also be taken into consideration.

At present no definitive statement can be made about the origin of the third-order depositional sequences observed in the non-marine post-rift sedimentary fill of the Pannonian basin. In order to better constrain these alternative explanations, this sedimentary succession should be described in terms of sequence stratigraphy (e.g. Vakarcs et al., in press), regardless of the obvious uncertainties about the mechanism governing their development.

B.2. THE SYN-RIFT SUCCESSION (MIDDLE MIOCENE)

The Pannonian Basin evolved in the Paratethys, which is characterized by its own faunal and tectonic evolution (for a summary, see Rögfl and Steininger, 1983); therefore regional stages were defined starting from the base of the Oligocene (see Fig. 3.4). Since the correlation of these regional stages with the standard ones is difficult due to the increasing endemism of the Paratethyan aquatic biota, sequence stratigraphy offers an excellent opportunity to correlate the Paratethyan stages with the Mediterranean ones (Tari et al., 1992a).

At first glance in the Neogene sedimentary succession of the Pannonian Basin, most of the unconformities defining the regional stages could be considered sequence boundaries (see Fig. 3.4) correlatable with the eustatic curves of Haq et al. (1987). Most of the Paratethyan stage boundaries are traditionally attributed to distinct orogenic phases. In contrast to the tectonically "quiet" post-rift phase dominated by thermal subsidence, one expects a complex interplay between eustasy and tectonics in the syn-rift phase. Indeed, some of the sequence boundaries seem to be tectonically enhanced sensu Vail et
al. (1991) as was suggested by Tari et al. (1992a). A nice example was published recently by Friebe (1993) who interpreted the approximately 16.5 Ma old Styrian unconformity as a tectonically enhanced sequence boundary in contrast to the traditional "orogenic" interpretation (e.g. Stille, 1924).

B.3. THE FLEXURAL SUCCESSION (MIDDLE EOCENE - EARLY MIOCENE)

Unlike the Middle to Late Miocene extensional evolution of the Pannonian Basin, the Late Paleogene basin evolution of Northern Hungary is best understood in terms of compressional tectonics (Chapter 5). Rapid basin subsidence caused by the load of backthrust inner Western Carpathian units resulted in a relatively deep, underfilled basin during Rupelian-Aquitanian times. In the Burdigalian, deposition took place principally in neritic water depths, indicating the cessation of thrusting and the beginning of "molasse" sedimentation. The Early Miocene evolution of the basin thus represents a major transgressive-regressive facies cycle, driven primarily by tectonics.

The third-order eustatic signal, however, is superimposed on this tectonic facies cycle. Based on detailed well log sequence stratigraphic analysis combined with sedimentological studies in outcrops, third-order depositional sequences were defined, which can be traced throughout the basin. Clear correlation can be found between these sequences and those indicated on the eustatic sea-level chart of Haq et al. (1987), within the resolution of available biostratigraphic data (Fig. B.2). Thus the pronounced change at the Kiscellian/Egerian (Rupelian/Chattian) boundary was related to the mid-Oligocene (30 Ma) eustatic sea-level fall of Haq et al. (1987) by Tari and Sztanó (1992). In this interpretation the Szöllőske Sandstone Member at the base of the Eger succession represents a submarine canyon fill deposited during the subsequent lowstand (slope fan systems tract). The glauconitic limestone unit of the Novaj Member is considered a condensed section associated with a maximum flooding surface (29 Ma). In the same
Fig. B.2. Late Oligocene - Early Miocene sequence stratigraphy in N Hungary (Tari and Sztanó, 1992).
manner, the younger Bretka Limestone unit (Fig. 1) shows the major early Miocene
eustatic transgression (24.8 Ma). The sudden inception of coarse clastic deposition
(Pétervására Sandstone) above siltstones deposited in significantly deeper water (Szécsény
"Schlier") is interpreted as the result of eustatic sea-level fall at the base of the Burdigalian
(21 Ma). Sedimentological studies in outcrops reveal the details of the sudden basinward
shift of various clastic depositional environments (Sztanó and Tari, 1993). Finally, at the
end of the Early Miocene, the isostatic uplift following the cessation of thrusting combined
with the increased amount of sediment supply (Budafok Sandstone) results in the
disappearance of marine sedimentation, in spite of the slow eustatic transgression (Fig.
B.2).

Although the third-order eustatic signal can be observed throughout the evolution of
the late Paleogene basin in Northern Hungary, a marked transgression at the base of the
Lower Oligocene Kiscell Clay (Hárshegy Sandstone, Fig. B.2) suggests that tectonics
alone is also capable of producing marked transgressional events of local character in
foredeep basins. This trangression at the base of the NP24 nannoplankton zone is
interpreted as the result of a major out-of-sequence thrusting event in the adjacent thrust-
fold belt (see Fig. 5.23), forming the tectonic load for a sudden basin subsidence phase. In
this particular case tectonics apparently overprinted eustasy (Tari and Sztanó, 1992).
APPENDIX C

LIST OF GEOLOGICAL MAPS USED IN THIS WORK

Bakony Mountains, geological map sheets, scale 1:20,000, Hung. Geol. Surv., Budapest.
Ajka 1976
Bakonybél 1978
Bakonycsernye 1982
Bakonyszentlászló 1983
Bakonyszentkirály 1983
Bakonyszombathely 1983
Bodajk 1983
Borzavár 1982
Csabrendek 1979
Csór 1982
Devecser 1979
Dudar 1981
Farkasgyepü 1978
Magyarpolány 1978
Mőr 1983
Öskü 1986
Pádragkút 1976
Sáska 1986
Súr 1980
Sümeg 1986
Szentgál 1976
Ugod 1978
Úrkút 1976
Veszprém 1980


Dumitrescu, I. and Sandulescu, M., 1970. Tectonic map of the Romanian Socialist Republic, scale: 1,000,000. Geologic Institute, Bucuresti.


Kilényi, E. and Sefara, J. (editors), 1989. Pre-Tertiary basement map of the Carpathian


APPENDIX D

LIST OF SEISMIC SECTIONS AND WELLS USED IN THIS THESIS

D.1 LIST OF SEISMIC SECTIONS AND WELLS OF REGIONAL SECTIONS

The list below summarizes all the seismic sections and wells that were used in the construction of regional sections in Chapter 3. Note, that in some cases wells were projected into the section from several km distance.

REGIONAL SECTION A
Seismic lines:
Zi-108, Sa-5, Zi-91, Zi-122, Zi-171, D-1, D-1/F, D-1/E.

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Vcsa-24, Vpe-35.

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REGионаl SECTION H
Seismic lines:

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2             Ke-5      Kerekegyháza 0.5 km SW
3             Jak-1     Jakabszállás 0.7 km NE
4             Org-3     Orgovány   0.6 km SW
5             Bug-2     Bugac       1.8 km NE
6             Pálm-3   Pálmomonostora 0.9 km NE
7             Szeged-11 Szeged      0.8 km NE
8             Uszi-1    Újszentiván 1.2 km NE

REGionaL SECTION I
Seismic lines:
Vje-91, Je-74, Je-64, Je-59, Xab-16/A, Xab-16, Xtf-1, Tl-1, Fa-32, Fa-52, Or-47, To-69.

Wells: Number    Code    Full name    Projection into section
1             Nks-1      Nagykökényes 5.6 km W
2             Tő-2       Tóalmás     on section
3             Fa-5       Farmos      1.7 km NE
4             Újszil-2   Újszilvás  1.6 km SW
5             Abony-1   Abony       2.0 km NE
6             Tősz-1    Tőszeg      on section
7             Nrév-1    Nagyréve    on section
8             Fáb-4     Fábiánsebestyén on section
9             Oros-3    Oroszáza    0.3 km SW
10            Oros-DNy-1 Oroszáza    on section
11            Pit-D-1   Pitvaros    on section

REGionaL SECTION J
Seismic lines:
Eg-12, Eg-18, Eg-21, Je-39/1, Je-39/2, Cag-2, Flg-77, Me-14, En-19, Ök-18, Gyu-30, Be-62, Be-105, Be-86, Be-76.
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De-72, De-55, De-65, De-23, De-37, Ny-1.

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**REGIONAL SECTION P**
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Koe-9, Ka-182, Ka-76, Ka-81, Ka-186, De-19, Ny-4.

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<td>Föl-5</td>
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<td>on section</td>
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<td>12</td>
<td>Föl-2</td>
<td>Földes</td>
<td>on section</td>
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<td>13</td>
<td>Sáránd-I</td>
<td>Sáránd</td>
<td>1.8 km NW</td>
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<td>14</td>
<td>Nyáb-1</td>
<td>Nyúrábrány</td>
<td>0.2 km N</td>
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**REGIONAL SECTION Q**
Seismic lines:

<table>
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<tr>
<th>Wells: Number</th>
<th>Code</th>
<th>Full name</th>
<th>Projection into section</th>
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<td>1</td>
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<td>Algyö-K-1</td>
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<td>on section</td>
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<tr>
<td>3</td>
<td>Hód-I</td>
<td>Hódmezővásárhely</td>
<td>1.5 km SE</td>
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<tr>
<td>4</td>
<td>Bés-1</td>
<td>Békéssámon</td>
<td>3.0 km SE</td>
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<td>5</td>
<td>Pf-17</td>
<td>Pusztáfoldvár</td>
<td>1.5 km SE</td>
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<td>6</td>
<td>Pf-32</td>
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<td>7</td>
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<tr>
<td>9</td>
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<td>Sarkad</td>
<td>on section</td>
</tr>
</tbody>
</table>


10 Komádi-DNY-1 Komádi on section
11 Komádi-12 Komádi on section
12 Sas-9 Mezősas on section
13 Bike-É-1 Bíharkeresztes 0.3 km NW
14 Kism-2 Kismarja 1.4 km E
15 Álm-4 Álmosd on section
16 Álm-12 Álmosd on section
17 Álm-K-1 Álmosd on section

REGIONAL SECTION R
Seismic lines:
He-24, Ma-25, Ma-32, Ma-20, To-60, To-69, To-58, Be-76, Be-84, Be-58.

Wells: Number Code Full name Projection into section
1 F-21 Ferencszállás 2.6 km SE
2 F-22 Ferencszállás 0.6 km SE
3 Makó Makó 4.0 km SE
4 Pit-1 Pitvaros 2.0 km SE
5 Mh-Ny-2 Mezőhegyes on section
6 Mh-Ny-1 Mezőhegyes on section
7 Vég-2 Végegházá 0.2 km NE
8 Kág-3 Kunágota on section

D.2 LIST OF SEISMIC SECTIONS IN THE STUDY AREA

<table>
<thead>
<tr>
<th>Area C (Csapod)</th>
<th>Area M (Mihályi)</th>
<th>Area P (Pásztori)</th>
<th>Area N (Nagyigmánd)</th>
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<td>M2 Vpa-104</td>
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<td>Area V (Vaszár)</td>
<td>Area D (Dabrony)</td>
<td>Area K (Káld)</td>
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<td>Area A (Andráshida)</td>
<td>Area S (Szentgotthárd)</td>
<td>Area Z (Zalatárnok)</td>
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<td>Z26 Zi-143</td>
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D.3 LIST OF SEISMIC SECTIONS AND WELL-LOGS USED IN THE STUDY AREA

Seismic lines:
(Geophysical Exploration Company)

Sa-12, 14, 15, 16, 19, 36, 39
Va-3, 4, 5, 17, 21, 22, 23, 24, 25, 26, 27, 28
Vcsa -1, 2, 4, 5, 6, 7, 8, 9, 10, 11, 12, 13, 14, 15, 16, 17, 18, 19, 20, 21, 23, 24, 25, 26, 27, 31, 32
Vik-9, 10, 11/1, 12
Vpá-7, 10, 14, 20, 23, 26, 27, 28, 29, 30, 31, 33, 36, 37, 38, 39, 40, 41, 42, 43, 44, 45, 46, 47, 48, 49/a, 50, 51, 52, 53, 54, 55, 56, 57, 58, 60, 63, 66/1, 66/2, 68, 69, 72, 73, 78, 79/1, 79/2 80, 81, 82, 83, 85, 89, 90, 91, 96, 98, 99, 100, 101, 102, 103, 104, 105, 106, 107, 108
Vpé-26, 27, 28, 29, 30, 40
Vva-1, 2, 3, 6, 7, 8, 9, 10, 11, 12 13, 14, 15, 16, 18, 19, 20
Zi-107, 108, 109, 110, 111

(Eötvös Geophysical Institute)

Wells with 1:500 scale SP and resistivity logs (Geophysical Exploration Company):

Ács-1
Bsz (Bakonyszücs) -1
Bor (Borgáta) -1
Bö (Bősárkány) -1
Bük-1 -2
CelI (Celldömölk) -1
Cell-ÉNy (Celldömölk) -1
Csapod-1
Da (Dabrony) -1
Gönyü-1
Gysz (Györszemere) -2 -3
Ike (Ikervár) -1 -2 -3 -4 -6 -8 -10 -11
Káld-1
Kár-1
Kol (Nemeskolta) -1 -2 -3
Mes (Mesteri) -1
M (Mihályi) -4 -10 -12 -15 -16 -17 -19 -21 -22 -28 -29 -32
MF (Mihály-felső) -4
Mos (Mosonszentjános) -1
Msz (Mosonzołnok) -1 -2
Nig (Nagyigmánd) -1 -3
Öl (Ölbö) -1 -3 -4 -6
Pá (Pásztori) -1 -2 -4
Pe (Pecöl) -1
Pér-1 -2
Pi (Pinnye) -1
Raj (Rajka) -1
Rád (Egyházasrádóc) -1 -2
Rás (Rábasömjén) -1
Sót (Sótony) -1 -2
Szany-1
Tak (Takácsi) -1 -2
Tét -1 -2 -3 -4 -5 -6
Ukk-1 -2
Va (Vasvár) -1
Vasz (Vaszar) -1 -2 -3 -4
VaszDNy (Vaszar) -1
Vát-1
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LEFT TO RIGHT, TOP TO BOTTOM, WITH SMALL OVERLAPS

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PLEASE NOTE:

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LEFT TO RIGHT, TOP TO BOTTOM, WITH SMALL OVERLAPS

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PANEL 2
REGIONAL SEISMIC SECTIONS
PANNONIAN BASIN
EASTERN HUNGARY
20 KM

WELL
TIE WITH OTHER REGIONAL LINE
SEISMIC TIE
MARKED CHANGE IN ORIENTATION

BOTTOM OF WELL:
M MIDDLE MIocene
V MIocene VOLCANICS
Pg PALEogene
F SZOLNOK FLYSCH
Mz MESozoIC
Pz PALEozoIC
Cr CRYSSTALLINE

LEGEND
TOP INTRAPANNONIAN SUBSEQUENCE
TOP INTRAPANNONIAN SEQUENCE
TOP MIDDLE MIocene (SARMATIAN)
TOP MIocene SYN-RIFT (OR VOLCANICS)
TOP PRE-MIocene (OR EARLY MIocene)
TOP SENONIAN (OR SZOLNOK FLYSCH)
TOP PRE-SENONIAN BASEMENT

INDEX MAP

NW SECTION
PLEASE NOTE:

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PLEASE NOTE:

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PANEL 4A
SEISMIC SECTIONS C1, C2 and C17
PANNONIAN BASIN NW HUNGARY
5 KM

LOCATION MAP

LIST OF SEISMIC ILLUSTRATIONS
- PANEL 4A: SEISMIC SECTIONS C1, C2 AND C17
- PANEL 4B: LINE DRAWINGS C1, C2 AND C17
- PANEL 5: LINE DRAWINGS C3, C4 AND C15
- PANEL 6: LINE DRAWINGS C5, C6 AND C18
- PANEL 7A: SEISMIC SECTIONS C7, C10 AND C20
- PANEL 7B: LINE DRAWINGS C7, C10 AND C20
- PANEL 8A: SEISMIC SECTIONS M1, M3 AND M18
- PANEL 8B: LINE DRAWINGS M1, M3 AND M18
- PANEL 9A: SEISMIC SECTIONS G2, G3 AND P26
- PANEL 9B: LINE DRAWINGS G2, G3 AND P26
- PANEL 10A: SEISMIC SECTIONS D9, D12 AND D31
- PANEL 10B: LINE DRAWINGS D9, D12 AND D31
- PANEL 11: LINE DRAWINGS D13, D14 AND D33
- PANEL 12: LINE DRAWINGS K3, K4 AND K11
PLEASE NOTE:

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LEFT TO RIGHT, TOP TO BOTTOM, WITH SMALL OVERLAPS

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PANEL 4B
SEISMIC SECTIONS C1, C2 and C17
PANNONIAN BASIN
NW HUNGARY
5 KM

LEGEND
POST-RIFT
TOP MIDDLE MIOCENE (SARMATIAN)
TOP MIOCENE SYN-RIFT (WITH CLASTIC FANS)
TOP UPPER AUSTROALPINE
TOP MIDDLE AUSTROALPINE
TOP LOWER (?) AUSTROALPINE
TOP PENNINIC

LOCATION MAP
20 KM

NW SECTION C1
0
1
2
3
4

NW SECTION C2
0
1
2
LIST OF SEISMIC ILLUSTRATIONS

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- PANEL 11  LINE DRAWINGS  D13, D14 AND D33
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- PANEL 13  LINE DRAWINGS  I1, I3 AND I16
- PANEL 14A SEISMIC SECTIONS  A5, A15 AND A16
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PANEL 5
SEISMIC SECTIONS C3, C4 and C15
PANNONIAN BASIN
NW HUNGARY
5 KM

LEGEND
POST-RIFT
TOP MIDDLE MIocene (SARMATIAN)
TOP MIOCENE SYN-RIFT (WITH CLASTIC FANS)
SYN-RIFT
TOP UPPER AUSTROALPINE
PRE-RIFT
TOP MIDDLE AUSTROALPINE
TOP LOWER(?) AUSTROALPINE
TOP PENNINIC

LOCATION MAP
20 KM

NW SECTION C3

NW SECTION C4
LIST OF SEISMIC ILLUSTRATIONS

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UMI
PANEL 6
SEISMIC SECTIONS C5, C6 and C18
PANNONIAN BASIN
NW HUNGARY
5 KM

LEGEND
POST-RIFT
TOP MIDDLE MIocene (Sarmatian)
TOP MIOCENE SYN-RIFT (WITH CLASTIC FANS)
TOP UPPER AUSTROALPINE
TOP MIDDLE AUSTROALPINE
TOP LOWER(?) AUSTROALPINE
TOP PENNINIC

LOCATION MAP
20 KM

NW SECTION C5

NW SECTION C6
LIST OF SEISMIC ILLUSTRATIONS

PANEL 4A  SEISMIC SECTIONS  C1, C2 AND C17
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PANEL 7A

SEISMIC SECTIONS C7, C10 and C20

PANNONIAN BASIN
NW HUNGARY

5 KM

LOCATION MAP

LIST OF SEISMIC ILLUSTRATIONS

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<td>8A</td>
<td>SEISMIC SECTIONS M1, M3 AND M18</td>
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</tr>
<tr>
<td>10B</td>
<td>LINE DRAWINGS D9, D12 AND D31</td>
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PANEL 7B
SEISMIC SECTIONS
C7, C10 and C20
PANNONIAN BASIN
NW HUNGARY

5 KM

LEGEND
POST-RIFT
TOP MIDDLE MIocene (GARMATIAN)
TOP MIOCENE SYN-RIFT (WITH CLASTIC FANS)
SYN-RIFT
TOP UPPER
AUSTROALPINE
PRE-RIFT
TOP MIDDLE
AUSTROALPINE
TOP LOWER(?)
AUSTROALPINE
TOP PENNINIC

LOCATION MAP
20 KM

NW SECTION C7
0
1
2
3
4

NW SECTION C10 C1
0
1
2

0
1
2
3
4
LIST OF SEISMIC ILLUSTRATIONS

PANEL 4A  SEISMIC SECTIONS  C1, C2 AND C17
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PANEL 6   LINE DRAWINGS  C5, C6 AND C18
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UMI
PANEL 8A
SEISMIC SECTIONS
M1, M3 and M18
PANNONIAN BASIN
NW HUNGARY
5 KM

LIST OF SEISMIC ILLUSTRATIONS
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NW SECT
0
1
2
3
4

NW SECT
0
1
2

LOCATION MAP
20 KM
PLEASE NOTE:

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PANEL 9A
SEISMIC SECTIONS
G2, G3 and P26
PANNONIAN BASIN
NW HUNGARY
5 KM

LOCATION MAP

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PANEL 15  LINE DRAWINGS  A8, A17 AND A20
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- PANEL 8B  LINE DRAWINGS M1, M3 AND M18
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PANNONIAN BASIN
NW HUNGARY

5 KM

LEGEND
POST-RIFT
TOP MIDDLE MIocene (SARMATIAN)
TOP MIocene SYN-RIFT (OR VOLCANICS)
SYN-RIFT
INTRA-SYNRIFT UNCONFORMITY
TOP MESOALPINE (PALEogene)
TOP EOALPINE (SENONIAN)
PRE-RIFT
INTRA-SENONIAN UNCONFORMITY
TOP UPPER AUSTRALPINE
EARLY ALPINE UNCONFORMITY

LOCATION MAP
20 KM

NW SECTION D9
M-20
M-22
LIST OF SEISMIC ILLUSTRATIONS

PANEL 4A   SEISMIC SECTIONS  C1, C2 AND C17
PANEL 4B   LINE DRAWINGS C1, C2 AND C17
PANEL 5    LINE DRAWINGS C3, C4 AND C15
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PANEL 8B   LINE DRAWINGS M1, M3 AND M18
PANEL 9A   SEISMIC SECTIONS  G2, G3 AND P26
PANEL 9B   LINE DRAWINGS G2, G3 AND P26
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PANEL 12   LINE DRAWINGS K3, K4 AND K11
PANEL 13   LINE DRAWINGS I1, I3 AND I16
PANEL 14A  SEISMIC SECTIONS  A6, A15 AND A16
PANEL 14B  LINE DRAWINGS A6, A15 AND A16
PANEL 15   LINE DRAWINGS A8, A17 AND A20
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PANEL 11
SEISMIC SECTIONS D13, D14 and D33
PANNONIAN BASIN NW HUNGARY
5 KM

LEGEND
POST-RIFT
TOP MIDDLE MIocene (SARMATIAN)
TOP MIocene SYN-RIFT (OR VOLCANICS)
SYN-RIFT
INTRA-SYNRIFFT UNCONFORMITY
TOP MESOALPINE (PALEogene)
TOP EOALPINE (SENONIAN)
INTRA-SENONIAN UNCONFORMITY
PRE-RIFT
TOP UPPER AUSTROALPINE
EARLY ALPINE UNCONFORMITY

LOCATION MAP

NW SECTION D13
LIST OF SEISMIC ILLUSTRATIONS

PANEL 4A  SEISMIC SECTIONS  C1, C2 AND C17
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PANEL 5   LINE DRAWINGS  C3, C4 AND C15
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PANEL 9A  SEISMIC SECTIONS  G2, G3 AND P26
PANEL 9B  LINE DRAWINGS  G2, G3 AND P26
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PANEL 10B LINE DRAWINGS  D9, D12 AND D31
PANEL 11  LINE DRAWINGS  D13, D14 AND D33
PANEL 12  LINE DRAWINGS  K3, K4 AND K11
PANEL 13  LINE DRAWINGS  I1, I3 AND I16
PANEL 14A SEISMIC SECTIONS  A6, A15 AND A16
PANEL 14B LINE DRAWINGS  A6, A15 AND A16
PANEL 15  LINE DRAWINGS  A8, A17 AND A20
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PANEL 12
SEISMIC SECTIONS K3, K4 and K11
PANNONIAN BASIN
NW HUNGARY
5 KM

LEGEND
POST-RIFT
TOP MIDDLE MIocene
(SARMATIAN)
TOP MIOCENE SYN-RIFT
(OR VOLCANICS)
SYN-RIFT
INTRA-SYNRIFT
UNCONFORMITY
TOP MesoALPINE
(PALEogene)
PRE-RIFT
TOP EOALpine
(SENONIAN)
INTRA-SENONIAN
UNCONFORMITY
TOP UPPER
AUSTROALPINE
EARLY ALPINE
UNCONFORMITY

LOCATION MAP
20 KM

NW SECTION K3
LIST OF SEISMIC ILLUSTRATIONS

PANEL 4A SEISMIC SECTIONS C1, C2 AND C17
PANEL 4B LINE DRAWINGS C1, C2 AND C17
PANEL 5 LINE DRAWINGS C3, C4 AND C15
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PANEL 7B LINE DRAWINGS C7, C10 AND C20
PANEL 8A SEISMIC SECTIONS M1, M3 AND M18
PANEL 8B LINE DRAWINGS M1, M3 AND M18
PANEL 9A SEISMIC SECTIONS G2, G3 AND P26
PANEL 9B LINE DRAWINGS G2, G3 AND P26
PANEL 10A SEISMIC SECTIONS D9, D12 AND D31
PANEL 10B LINE DRAWINGS D9, D12 AND D31
PANEL 11 LINE DRAWINGS D13, D14 AND D33
PANEL 12 LINE DRAWINGS K3, K4 AND K11
PANEL 13 LINE DRAWINGS I1, I3 AND I16
PANEL 14A SEISMIC SECTIONS A6, A15 AND A16
PANEL 14B LINE DRAWINGS A6, A15 AND A16
PANEL 15 LINE DRAWINGS A8, A17 AND A20
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PANEL 13
SEISMIC SECTIONS I1, I13 and I16
PANNONIAN BASIN
NW HUNGARY
5 KM

LEGEND
POST-RIFT
TOP MIDDLE MIocene
(SARMATIAN)
TOP MIocene SYN-RIFT
(OR VOLCANICS)
SYN-RIFT
INTRA-SYNRIFT
UNCONFORMITY
TOP MESOALPINE
(PALEogene)
TOP EOALPINE
(SENONIAN)
PRE-RIFT
INTRA-SENONIAN
UNCONFORMITY
TOP UPPER
AUSTRALPINE
EARLY ALPINE
UNCONFORMITY

LOCATION MAP
20 KM

NW SECTION I1
LIST OF SEISMIC ILLUSTRATIONS

PANEL 4A  SEISMIC SECTIONS C1, C2 AND C17
PANEL 4B  LINE DRAWINGS C1, C2 AND C17
PANEL 5   LINE DRAWINGS C3, C4 AND C15
PANEL 6   LINE DRAWINGS C5, C6 AND C18
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PANEL 14B
SEISMIC SECTIONS A6, A15 and A16
PANNONIAN BASIN
NW HUNGARY
5 KM

LEGEND
POST-RIFT
TOP MIDDLE MIOCENE (SARMATIAN)
TOP MIOCENE SYN-RIFT (OR VOLCANICS)
SYN-RIFT
INTRA-SYNRIFT UNCONFORMITY
TOP MESOALPINE (PALEogene)
TOP EOALPINE (SERNONIAN)
INTRA-SEALPINE UNCONFORMITY
PRE-RIFT
TOP UPPER AUSTROALPINE
EARLY ALPINE UNCONFORMITY

LOCATION MAP

SW SEC
**LIST OF SEISMIC ILLUSTRATIONS**

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PANEL 15
SEISMIC SECTIONS
A8, A17 and A20
PANNONIAN BASIN
NW HUNGARY
5 KM

LEGEND
POST-RIFT
TOP MIDDLE MIocene (Sarmatian)
TOP MIOCENE SYN-RIFT
(OR VOLCANICS)
SYN-RIFT
INTRA-SYNRIFT UNCONFORMITY
TOP MESOALPINE
(PALEogene)
TOP EOALPINE
(SENONIAN)
INTRA-SENONIAN
UNCONFORMITY
PRE-RIFT
TOP UPPER
AUSTROALPINE
EARLY ALPINE
UNCONFORMITY

LOCATION MAP
20 KM

SW SECTION A17
0 NL-228 Ho-2
1
2
LIST OF SEISMIC ILLUSTRATIONS

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PLATE 1
SEISMIC SECTION Ca (K-4/88)
PANNONIAN BASIN
WESTERN HUNGARY
10 KM

LEGEND

POST-RIFT

TOP MIDDLE MIocene
(Sarmatian)

TOP MIocene SYN-RIFT
(WITH CLASTIC FANS)

TOP UPPER
AUSTROALPINE

SYN-RIFT

TOP MIDDLE
AUSTROALPINE

PRE-RIFT

TOP LOWER(?)
AUSTROALPINE

TOP PENNINIC

INDEX MAP

100 KM
PLEASE NOTE:

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PLATE 3
SEISMIC SECTION Da (MK-1/80-82)
PANNONIAN BASIN
WESTERN HUNGARY
10 KM

LEGEND
- POST-RIFT
- TOP MIDDLE MIOCENE (SARMATIAN)
- TOP MIOCENE SYN-RIFT (WITH CLASTIC FANS)
- TOP UPPER AUSTROALPINE
- TOP MIDDLE AUSTROALPINE
- TOP LOWER (?) AUSTROALPINE
- TOP PENNICIC

INDEX MAP

100 KM

NW SOPRON HIGH

FER
PLEASE NOTE:

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PLATE 4
SEISMIC SECTION Dd (MK-1/82)
PANNONIAN BASIN
WESTERN HUNGARY
5 KM

LEGEND

POST-RIFT
- TOP MIDDLE MIOCENE
  (SARMATIAN)
- TOP MIOCENE SYN-RIFT
  (OR VOLCANICS)

SYN-RIFT
- INTRA-SYNRFIT
  UNCONFORMITY
- TOP MESOALPINE
  (PALEogene)
- TOP EOALPINE
  (SENONIAN)

PRE-RIFT
- INTRA-SENONIAN
  UNCONFORMITY
- TOP UPPER
  AUSTROALPINE
- EARLY ALPINE
  UNCONFORMITY

INDEX MAP

- GOOD CONDUCTIVITY LAYER
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UMI
PLATE 6
SEISMIC SECTION Fa (MK-3/77)
PANNONIAN BASIN
WESTERN HUNGARY
10 KM

LEGEND
POST-RIFT
TOP MIDDLE MIOCENE (SARMATIAN)
TOP MIOCENE SYN-RIFT (OR VOLCANICS)
SYN-RIFT
INTRA-SYNRIFT UNCONFORMITY
TOP Mesoalpine (PALEogene)
TOP EOALPINE (SENONIAN)
PRE-RIFT
INTRA-SENONIAN UNCONFORMITY
TOP UPPER AUSTROALPINE
EARLY ALPINE UNCONFORMITY

INDEX MAP

100 KM

NW DANUBE

TWO WAY TIME IN SECONDS
0 2 4 6 8 10

NW DANUBE

TWO WAY TIME IN SECONDS
0 2 4 6 8 10
PLEASE NOTE:

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PLATE 7
SEISMIC SECTION Jb (PGT-1)
PANNONIAN BASIN
EASTERN HUNGARY
20 KM

INDEX MAP

LEGEND

TOP INTRAPANNONIAN SUBSEQUENCE

TOP INTRAPANNONIAN SEQUENCE

TOP MIDDLE MIocene (SARMATIAN)

TOP MIocene SYN-RIFT (OR VOLCANICS)

TOP PRE-MIOCENE (OR EARLY MIocene)

TOP SENONIAN (OR SZOLNOK FLYSCH)

TOP PRE-SENONIAN BASEMENT
PLEASE NOTE:

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PLATE 9
SEISMIC SECTIONS Z20, Z21
PANNONIAN BASIN
NW HUNGARY
5 KM

LEGEND
- TOP INTRAPANNONIAN SUBSEQUENCE
- TOP INTRAPANNONIAN SEQUENCE
- TOP MIDDLE MIocene (SARMATIAN)
- TOP MIocene SYN-RIFT (OR VOLCANICS)
- TOP PRE-MIocene (OR EARLY MIocene)
- TOP SENONIAN (OR SZOLNOK FLYSCH)
- TOP PRE-SENONIAN BASEMENT

INDEX MAP

100 KM
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PLATE 10
REGIONAL STRUCTURE SECTION
NW PANNONIAN BASIN
AUSTRIA/HUNGARY
SCALE: 1:200,000

NO VERTICAL EXAGGERATION
EASTERN ALPS

LASSE BASIN  RHENODANUBIAN FLYSCH ZONE  NORTHERN CALCAREOUS ALPS BAJUVARICUM  TIROLICUM

SECTION OF WESSELY (1988)

OLTENBACH-1  MANZING-1  BERNDORF-1

EUROPEAN FORELAND
MIOCENE-RECENT
MOLASSE SEDIMENTS
CRETACEOUS-PALEogene
FLYSCH SEDIMENTS
MESozoIC PASSive
MARGIN SEDIMENTS
PALEozoIC HERCYNIAN
CRYSTALLINE ROCKS

AUSTROALPINE NAPPE SYSTEM AND TERRAIN

UPPER MIOCENE-RECENT
POST-RIFT SEDIMENTS
MIDDLE MIOCENE
SYN-RIFT SEDIMENTS
PALEogene SEDIMENTS
UPPER CRETACEOUS "GOSAU" SEDIMENTS

Mesozoic Austroal
Paleozoic Austroal
Paleozoic Lower Austrian
Mesozoic Penninic

• TOP OF PRE-TERTIARY BASEMENT DETERMINED BY MAGNETOTELLURICS
• GOOD CONDUCTIVITY LAYER DETERMINED BY MAGNETOTELLURICS
STEM AND TERTIARY BASINS

- MESOZOIC UPPER
  - AUSTROALPINE UNITS
- PALEOZOIC UPPER
  - AUSTROALPINE UNITS
- PALEOZOIC MIDDLE AND LOWER AUSTROALPINE UNITS
- MESOZOIC
  - PENNINIC UNITS

MOHO?