## High-resolution morpho-tectonic profiling across an orogen


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# 4 Tectonic and climatic controls on river incision and terrace formation

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## 4.1. Introduction

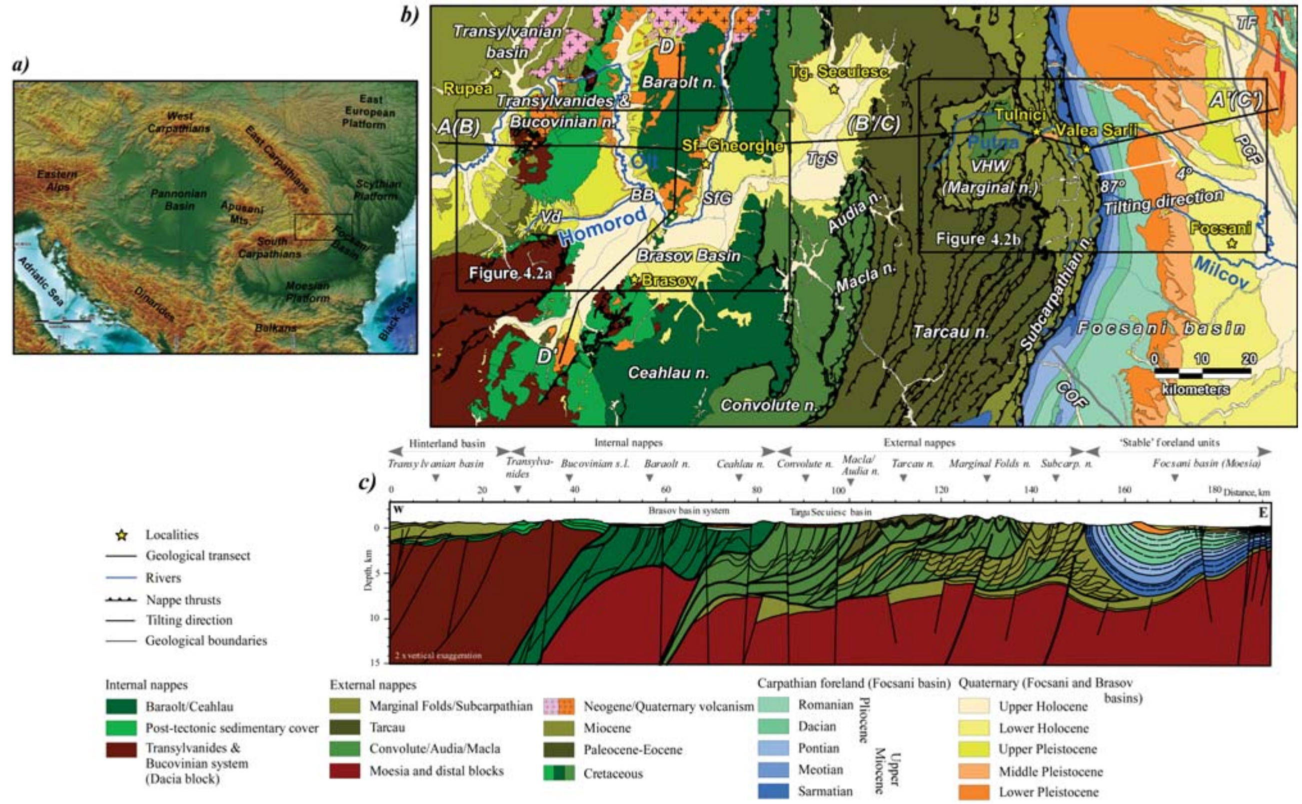
Fluvial activity in an active mountain chain results from interaction between rock uplift and erosion. Tectonic compressional processes tend to build up topography, while the surface ones (fluvial, glacial and hillslope) tend to tear them down (Burbank and Anderson, 2001). In order to distinguish between the dominant processes controlling fluvial incision, it is essential to understand the river behaviour over time and its response to long-term events/changes, which is recorded as river terraces. Rates and amounts of river incision plotted against geological cross-section and longitudinal river profiles helped also this research to constrain the relationship between surface processes and behaviour of underlying bedrock (Burbank et al., 1996).

This type of active processes can be studied in detail in the SE Carpathians, where a wealth of recent upper crustal information exists and can be connected with the recent post-collisional evolution of this orogen (Matenco et al., 2007 and references therein). A key element in discriminating between the different structural models of the area at shallow crustal levels is the rate of vertical movements during the Pliocene-Quaternary inversion of the Pannonian-Carpathian system (Cloetingh et al., 2006), which proved to be near the resolution of the structural analysis in the SE Carpathians (Leever et al., 2006).

This study reconstructs the Middle Pleistocene-Holocene morphotectonic and paleoclimatic evolution by analysing the amplitude and timing of vertical movements affecting the orogenic system along a 175 km-long cross-section. It starts in the Miocene sediments of the Transylvanian hinterland basin and ends in the flat-lying Holocene deposits of the Focşani foreland (Fig. 4.1b). The Quaternary uplift is obvious in structural observations and from the ~9°ENE-ward tilting of the Lowermost Pleistocene strata (Necea et al., 2005). This is also shown by the subsequent uplift of the river terraces at different elevations above the riverbed level, equally distributed over the internal and external orogenic nappes and the hinterland and foreland basin (Fig. 4.2a-b).

The foreland river system (e.g. Putna and Milcov rivers) runs eastward, cutting transversally the frontal thrusts of the external Carpathian nappes and the Uppermost Miocene-Quaternary strata on the western flank of the Focşani basin (Fig. 4.2a). In the hinterland, the river network (e.g. Olt and Homorod rivers) runs parallel to the internal orogenic nappes and the Braşov intramontane basin (Fig. 4.2b). The Middle Pleistocene-Holocene uplift of the SE Carpathians favoured a deep incision of this river network in the east, leaving numerous terrace levels at different elevations, e.g. 250 m from the riverbed (Figs. 4.3a and 4.4a). This maximum recorded uplift is located in the centre of the orogen, gradually decreasing east- and westward towards the foreland and the hinterland, respectively, only 100 m of uplift being recorded in the latter (Figs. 4.3b and 4.4b).

Figure 4.1




Digital elevation model of the Eastern Alps-Carpathians-Dinarides-Balkans region, with the study area located in the northern part of the SE Carpathians bending zone. (b) Detailed geological map with the W-E geological transect (A-A' profile) along which amplitudes and rates of vertical movements have been quantified. The two insets are areas where detailed geomorphological studies were performed, namely the orogen-foreland transition (E) and the intramontane basin (W), respectively (detailed in Fig. 4.2a-b). Note the Putna and Milcov rivers drain transversally the external orogenic units and the Focşani foreland basin, whereas the Olt and Homorod rivers flow parallel to the internal units and the Braşov intramontane basin. D-D' is the topographic profile shown in Fig. 4.11. Vd-Vlădeni, BB-Bârsa-Baraolt and SfG-Sfântu Gheorghe sub-basins and TgS-Târgu Secuiesc basin; VHW-Vrancea half-window; TF-Trotuş fault, PCF-Peceneaga-Camena fault and COF-Capidava-Ovidiu fault. (c) Regional geological cross-section along which the vertical movements were quantified (modified after Leever et al., 2006; Schmid et al., 2008).

The time when the river terraces formed are crucial in determination of river incision rates and represent the target of this study. It should be noted that the resolution of geomorphological analyses and biostratigraphic age constraints is limited (Ghenea et al., 1971; Posea, 1981). In order to improve the resolution, this research dates loess sequences covering a wide range of erosional surfaces by means of infrared stimulated luminescence (IRSL) dating method. Subsequently, these ages were correlated with Marine Isotope Stages (MIS) and chrono-stratigraphy of the Alps-Carpathians realm. Few terrace levels could not be dated with IRSL (i.e. Braşov intramontane basin) and for those levels river incision rates were determined by using previously published ages (e.g. Fig. 4.9). Additionally, the river incision rate obtained for an older terrace was used to derive the formation ages for younger levels (i.e. Focşani basin; Fig. 4.9 and Table 4.3a-b).

Lastly, this study discusses some of the possible underlying mechanism(s) for the uplift. To this end, obtained river incision rates are plotted on the general structure of the SE Carpathians, considering the overall relationship between the surface processes and the bedrock behaviour. This serves to reconstruct the evolution of the SE Carpathians during the Middle Pleistocene-Holocene by means of vertical movements and mechanisms for uplift/erosion.

## 4.2. Neogene and Quaternary geological setting of the SE Carpathians

The Carpathian orogenic system (Fig. 4.1a) formed in response to the Middle Triassic-Tertiary evolution of various continental blocks. Three of these blocks are presently located in the internal part of the orogen, i.e. Tisza to the south-west, Dacia to the south-east and ALCAPA to the north of the orogen (Csontos and Vörös, 2004). The orogen is surrounded to the exterior by the East European/Scythian/Moesian platforms to the north, east and south (Visarion et al., 1988). During the Middle Triassic-Tertiary , the blocks were separated by two oceanic domains, namely the Transylvanides to the west and south and the Ceahlău-Severin to the east and north, respectively (Săndulescu, 1984).

The Cretaceous evolution of the SE Carpathians is linked with the closure of the Transylvanides domain, when contraction and continental collision occurred between the Tisza and Dacia blocks in intra-Albian time. This is also responsible for the internal deformation of the Bucovinian nappes and their eastward thrusting over the more external Ceahlău-Severin nappes (Săndulescu, 1988). The Paleogene is regarded as a large period of quiescence, when the orogen was apparently inactive for ~40 Ma (Săndulescu, 1988; Ştefănescu et al., 1988). The Neogene evolution of the SE Carpathians is linked with the final closure of the Ceahlău-Severin ocean and progressive deformation of its eastern passive continental margin, which was entirely subducted by the moment when the Carpathian nappe emplacement ceased at ~11 Ma (Matenco and Bertotti, 2000 and references therein).

The gradual exhumation of the East Carpathians appears laterally variable along the strike of the orogen, e.g. the Middle-Late Miocene collisional related-exhumation affecting its central part was replaced by the post-collisional unroofing in the SE bending area, postdating the main collisional event (11 Ma; Sanders et al., 1999).

### 4.2.1. Focşani foreland basin

The rather unusual 13 km-thick Middle Miocene-Quaternary sediments of the Focşani basin (Tărăpoancă et al., 2003 and references therein) are the result of spatial juxtaposition of several mechanisms (Matenco et al., 2007). Initially, the Focşani basin was affected by a Middle Miocene phase of extension. Secondly, flexural subsidence took place in the area when it became part of the Late Miocene foredeep basin of the East Carpathians. The associated sediments were moreover covered by those coeval with the subsequent Latest Miocene-Pliocene generalised subsidence, which affected the entire external part of the already locked SE and South Carpathians (Matenco et al., 2007).

However, the required accommodation space for the large sedimentary thickness is commonly explained by a slab-pull mechanism (senso Royden, 1993), as evidenced from observations of intermediate mantle, high-velocity anomaly observed in the SE Carpathians (Martin et al., 2006 and references therein). The Focşani basin is situated east of this anomaly. Subsequently, the Focşani basin was affected by post-collisional compressional deformation that was strictly localized in the SE Carpathians and peaked somewhere near the limit between the Pliocene and Quaternary. This is recorded through the uplift of the orogenic wedge and subsidence in its foreland, the resulting geometry being the syncline-like shape of the Focşani basin (Fig. 4.1c).

Sedimentary deposits on the western flank of the Focşani basin are strongly tilted eastward (Fig. 4.1b-c), with a near-vertical position at the contact with the orogen, the dip decreasing progressively towards the basin centre, reaching 15° in the Upper Pliocene strata. The source area of these deposits indicates a dominant southward transport direction, along the strike of the orogen (Negulescu, 2001), possibly as a result of the enhanced exhumation recorded in the central part of the East Carpathians (Sanders et al., 1999). The first major recognizable uplift of the external orogenic part and the western Focşani flank took place during the Late Pliocene. This is demonstrated by the overlying angular unconformity, the associated syn-tectonic sediments having a wedge-type geometry closer to the foreland (Fig. 4.1b, Necea et al., 2005; Leever et al., 2006). The subsequent Earliest Pleistocene subsidence recorded in the Focşani basin was characterized by rapid deposition of 400-500 m lacustrine sands and gravels with source areas located in the external Carpathian nappes and the western flank of the Focşani basin.

Two periods of increased uplift took place during the late Early Pleistocene and the late Middle Pleistocene-Holocene, respectively. These deformations resulted in the uplift of the western Focşani basin flank and tilted the entire structure ~9°ENE-ward (Necea et al., 2005).

During the first period tilting resulted in ~750 m uplift of the Lowermost Pleistocene strata above the riverbeds. By simple linear extrapolation of the tilted beds to the west, 3000 m or higher uplift can be estimated in the Marginal Folds nappe (Fig. 4.10), which fits well with the published AFT data to the north and south of the study area. Because the biostratigraphical age of the Lowermost Pleistocene deposits is poorly constrained by the paleontological data, the onset of the late Early Pleistocene uplift was tentatively placed around 1.3 Ma, middle of Early Pleistocene (1.81-0.781 Ma, Gradstein et al., 2004). These general literature time constraints gave a tentative uplift rate of 1.5 mm/yr for the time interval of 0.5 Ma and an uplift of 750 m.

The second late Middle Pleistocene-Holocene period achieved a cumulative uplift of ~250 m above the riverbeds (Chapter 2; Necea et al., 2005). This second period of uplift forms the focus of this study, providing higher resolution on time constraints using the IRSL method.

### 4.2.2. Braşov intramontane basin

In the hinterland, the subsidence of the Transylvanian basin led to deposition of up to 3.5 km of Middle-Upper Miocene sediments, this basin being completely exhumed during the Latest Pannonian, ~9 Ma (Krézsek and Bally, 2006 and references therein). This episode largely predates the onset of inversion that took place in the Late Pliocene and therefore cannot be responsible for the recent enhanced exhumation and erosion of the SE Carpathians.

The Pliocene extension, affecting the internal orogenic nappes, led to the formation of the Braşov intramontane basin system, which is geomorphologically made up of four sub-basins (Vlădeni, Bârsa-Baraolt, Sfântu Gheorghe and Târgu Secuiesc; Figs. 4.1b-c and 4.3b). This event is usually interpreted as resulting from strike-slip fault bounded extension, as evidenced by numerous N-S to NE-SW-trending normal faults that are mapped in outcrop-scale kinematic studies (Chalot-Prat and Gîrbacea, 2000; Fielitz and Seghedi, 2005 and references therein). This corroborates with the earlier interpretations of Ştefănescu et al. (1988) and is considered to be coeval with the Pliocene-Quaternary alkaline magmatism (Seghedi et al., 2004).

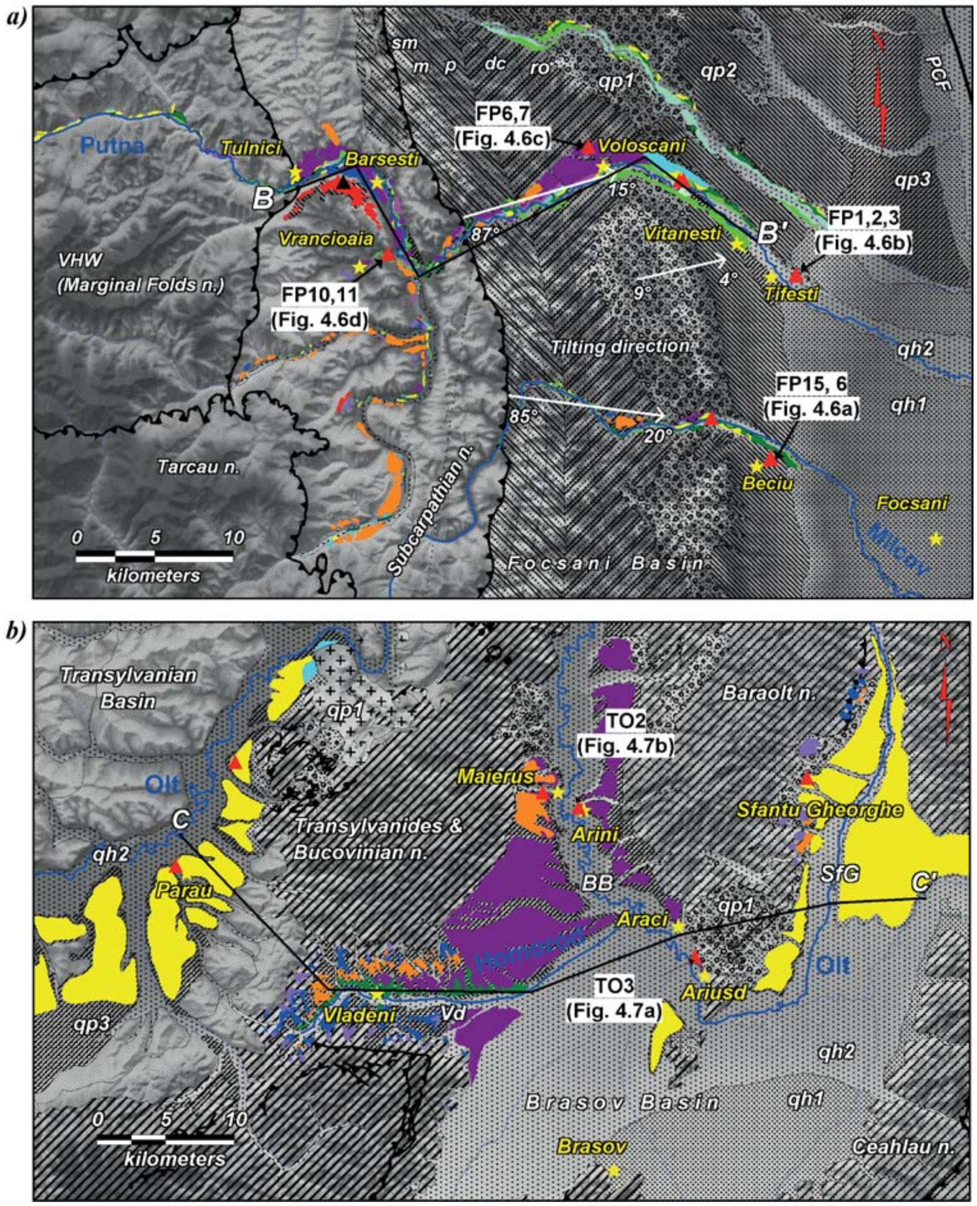
However, whether the subsidence is controlled by normal faulting is not fully clear. Many of the observed faults have more recently been interpreted as shallow volcanic collapse structures (Szakacs and Krézsek, 2006). Furthermore, part of the structures that presumably controlled the subsidence are either missing or do not align to the basin trend (i.e. faults are E-W-oriented) (Posea, 1981). Whether or not, significant normal faults controlled the subsidence, the Braşov basin provided space for for accumulation of volcanoclastic and lacustrine sediments, interbedded with coal deposits (Ghenea, 1970).

Interestingly, these sediments contain a fauna assemblage which is similar with the endemic one recorded eastward in the Carpathian foreland, e.g. Eastern Paratethys type (Rögl, 1999; Marinescu and Popescu, 1995; Olteanu, 2003). This requires an open connection between these basins across the present-day orogen. Such connection has been maintained until the Early Pleistocene as shown by the mollusca fauna which still points to lacustrine sedimentation of 50-100 m thick deposits (Villafranchian; Saulea et al., 1968; Ghenea et al., 1971). The Pliocene-Early Pleistocene volcano-sedimentary succession has variable thicknesses, ranging from 150 m in the proximal area to more than 500 m in the centre of (sub-) basin(s) (Ghenea et al., 1971; Marinescu et al., 1981; Laszlo, 2005).

The onset of inversion is recorded during the late Early Pleistocene in the Braşov basin, when lacustrine was replaced by fluvial sedimentation, being well observed during the Middle Pleistocene (Mindelian glacial) and the Late Pleistocene (Rissian-Würmian interglacial; Ghenea et. al., 1971). The river network has incised into the basin filling and underlying internal Carpathian nappes, resulting in formation of few levels of river terraces. The exact moment of inversion is rather poorly constrained, probably taking place during the late Early Pleistocene (1.3 Ma) similar as in the foreland basin. This uncertainty might bring errors when calculating the uplift rates (Subchapter 4.6).

The differential tectonic uplift and basin inversion recorded in the Braşov basin and the SE Carpathians were qualitatively estimated by means of geomorphology (Necea et al., 2005), structural studies (Ghenea et al., 1971; Posea, 1981; Rădoane et al., 2003; Fielitz and Seghedi, 2005) and validated by modern GPS studies, although the latter still lack in sufficient vertical resolution due to short-term measurements (van der Hoeven et al., 2005; Schmitt et al., 2007). In more details, the enhanced uplift located eastward has induced a lateral shift of the river network in the Braşov basin, which changed its initial foreland-ward to the present-day westward flow direction (Fielitz and Seghedi, 2005).

Figure 4.2




**Middle Pleistocene-Holocene river terraces in the SE Carpathians. (a) Eleven levels previously mapped between the Tulnici and Vităneşti localities in the external nappes and Focşani foreland basin (Necea et al., 2005). Note the highest/oldest level placed at ~250 m above the actual riverbed and the wide-spread level T5 of 39-41 m. B-B' profile corresponds to Putna river terraces (Fig. 4.3a). FP corresponds to luminescence sampling sites further detailed in Fig. 4.6a-d. VHW-Vrancea half-window, PCF-Peceneaga-Camena fault. (b) Six terrace levels occur between the Pârău and Sfântu Gheorghe localities, overlapping the internal nappes, the Braşov intramontane basin and the SE part of the Transylvanian hinterland basin. Note the oldest level placed at 100 m above the actual riverbed. C-C' profile corresponds to Olt river terraces (Fig. 4.3b; Ghenea et a., 1971 and this study). TO corresponds to luminescence sampling sites further detailed in Fig. 4.7a-b. Vd-Vlădeni, BB-Bârsa-Baraolt and SfG-Sfântu Gheorghe sub-basins.**

**Figure 4.2 (legend)**

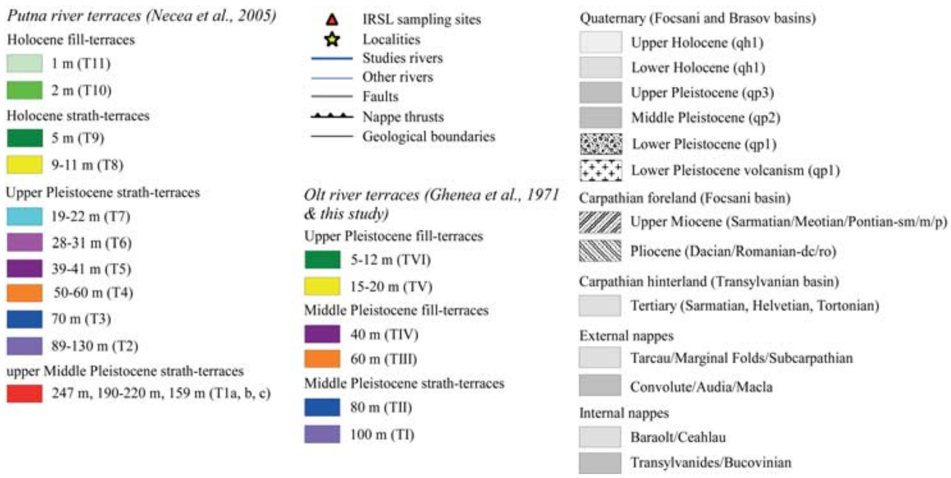



**Table 4.1**

***a)***

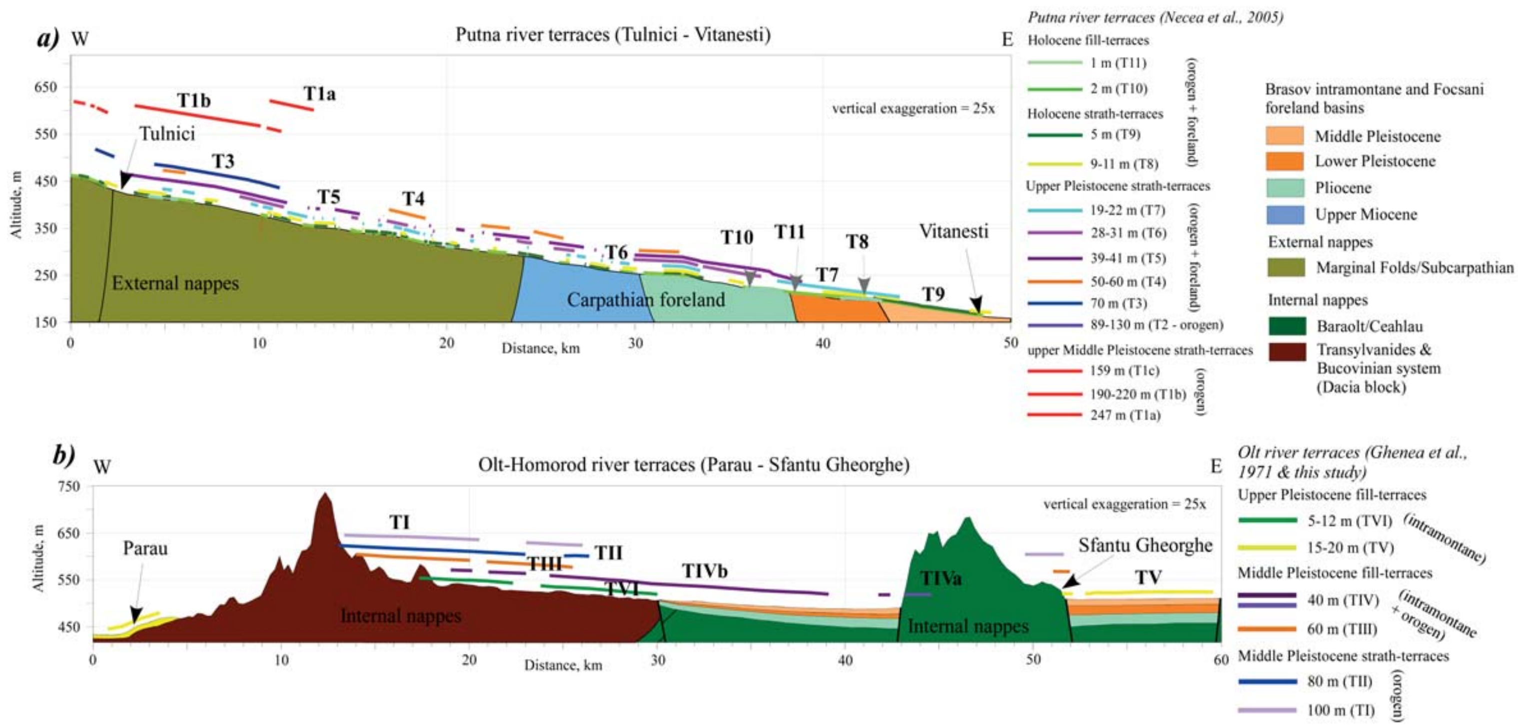
| Terrace level and regional loess seq./ RLS (occurrence) | Type of terrace or other features | Elevation of topo. surface above riverbed (m) | Thickness of sedim. cover (m) | Elev. of bedrock or denudation surface above riverbed (m) | Data source for terrace age N/D/T | Reference for terrace age |
|---|---|---|---|---|---|---|
| RLS1 (foreland) | Reg. loess seq. 1 | 67 | 18 | 49 | N | This study |
| T1a (orogen) | Strath-terrace | 247 | 3-7 | 237-242/239" | D and T | Grumazescu 1973 |
| T1b (orogen) | Strath-terrace | 190-220 | 3-7 | 187-210/198" | D | This study |
| T1c (orogen) | Strath-terrace | 159 | 3-7 | 149-154/151" | D | This study |
| RLS2 (foreland) | Reg. loess seq. 2 | 40 | 10 | 30 | N | This study |
| T2 (orogen) | Strath-terrace | 89-130 | 3-9 | 87-116/101" | D | This study |
| T3 (orogen+foreland) | Strath-terrace | 70 | 3-9 | 61 | D | This study |
| T4 (orogen+foreland) | Strath-terrace | 51-60 | 3-9 | 46-53/49" | D | This study |
| T5 (foreland) | Strath-terrace | 39-41 | 6 | 35-40/37" | N | This study |
| (orogen) | | 39-41 | 6 | 35-40/37" | N | This study |
| T6 (orogen+foreland) | Strath-terrace | 28-31 | 8 | 22 | D | This study |
| T7 (orogen+foreland) | Strath-terrace | 19-21 | 2-4 | 15 | D | This study |
| T8 (orogen+foreland) | Strath-terrace | 9-11 | 2-4 | 7 | D | This study |
| T9 (orogen+foreland) | Strath-terrace | 5 | 2-4 | 2.5 | D | This study |
| T10 (orogen+foreland) | Fill-terrace | 2 | - | - | T | Grumazescu 1973 |
| T11 (orogen+foreland) | Fill-terrace | 1 | - | - | T | Grumazescu 1973 |

***b)***

| Terrace level and regional loess seq./ RLS (occurrence) | Type of terrace or other features | Elevation of topo. surface above riverbed (m) | Thickness of sedim. cover (m) | Elev. of bedrock or denudation surface above riverbed (m) | Data source for terrace age N/D/T | Reference for terrace age |
|---|---|---|---|---|---|---|
| TI (orogen) | Strath-terrace | 100 | 8-10 | 90 | T | Ghenea et al., 1971 |
| TII (orogen) | Strath-terrace | 80 | 5-10 | 70 | T | Ghenea et al., 1971 |
| TIII (intramontane) | Fill-terrace | 60-65 | 10-15 | 50 | T | Ghenea et al., 1971 |
| TIVa (orogen) | Strath-terrace | 26 | 6 | 20 | N | This study |
| TIVb (intramontane) | Fill-terrace | 30-40 | 10 | 28 | N | This study |
| TV (intramontane) | Fill-terrace | 15-20 | - | ~17 | T | Ghenea et al., 1971 |
| TVI (intramontane) | Fill-terrace | 5-12 | - | ~6 | T | Ghenea et al., 1971 |

**Chronology and elevations of river terraces and regional loess sequences, and age determination methods (new, derived and traditional) for the two studied areas: (a) at the transition zone from the Carpathian orogen to the Focşani basin and (b) in the Braşov intramontane basin. Elevations of river terrace surfaces, terrace bedrocks and denudation surfaces are the relative vertical distances calculated between each of these surfaces and the actual riverbed. Data sources for terrace ages are: N-new IRSL ages obtained during this study by dating the loess sequences covering the terrace bedrocks and denudation surfaces, D-derived ages resulting from extrapolation of a river incision rate obtained at a higher terrace level and T-traditional data taken from the published literature, e.g. Grumăzescu (1973) for the Focşani basin and Ghenea et al. (1971) for the Braşov basin (explanation in text). 239'' are the mean elevation values for some bedrock surfaces used to calculate the river incision/uplift rate.**

Figure 4.3




**Chronology and altitudes of river terraces along the Putna (a) and Olt-Homorod (b) river profiles. Note in (b) that terraces are plotted along the Olt and Homorod rivers because the Olt river flows parallel to the orogen strike contradicting with the W-E Putna profile.**

## 4.3. Quantification of geomorphological data

### 4.3.1. Bedrock elevations, chrono-stratigraphy and ages of river terraces (new, derived and traditional data)

The factors used as principal data for deriving the amplitude of the vertical movements and for calculating the river incision rate are firstly the elevations of terrace bedrocks and denudation surfaces and secondly the timing of their formation .

Along with recognizing the elevations of terrace and denudation surfaces comes the differentiation of terrace types. Two types of river terraces are discussed in this study: strath- and fill-terraces (see Chapter 2; Burbank and Anderson, 2001). A strath-terrace is an erosional surface of the bedrock and overlain by a sedimentary fluvial cover of variable thickness (Table 4.1a-b), which may be covered by an additional loess sequences. Fill-terraces instead are erosional surfaces developed in sedimentary (fluvial) filling.

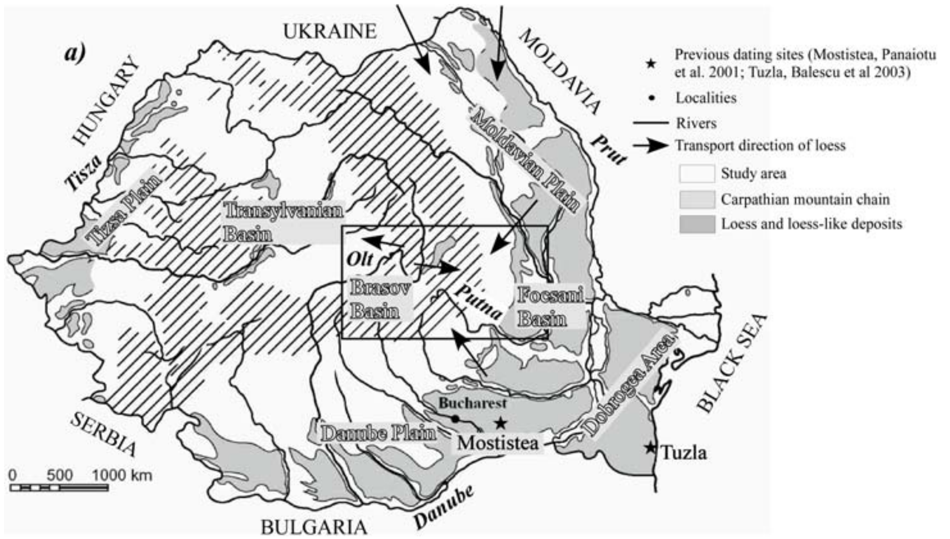
Elevations of the terrace erosional surfaces in the eastern part of the study area (i.e. Focşani basin; right rectangle in Fig. 4.1b) are mainly taken from the previous discussed geomorphological study carried out in the external nappes and the western flank of the Focşani basin (i.e. Chapter 2; Necea et al., 2005). The thicknesses of the fluvial deposits covering these surfaces are mainly taken from a study of Grumăzescu (1973; Table 4.1a). Eleven terrace levels have been mapped in this area (levels T1-T11; Fig. 4.2a), being located at various elevations above the riverbed, as high as 250 m in the orogenic nappes to 1-2 m in the foreland (Fig. 4.3a). These are mostly strath- and partly fill-terraces (Table 4.1.a). Denudation surfaces are present within the upper deposits of the monoclinal structure.

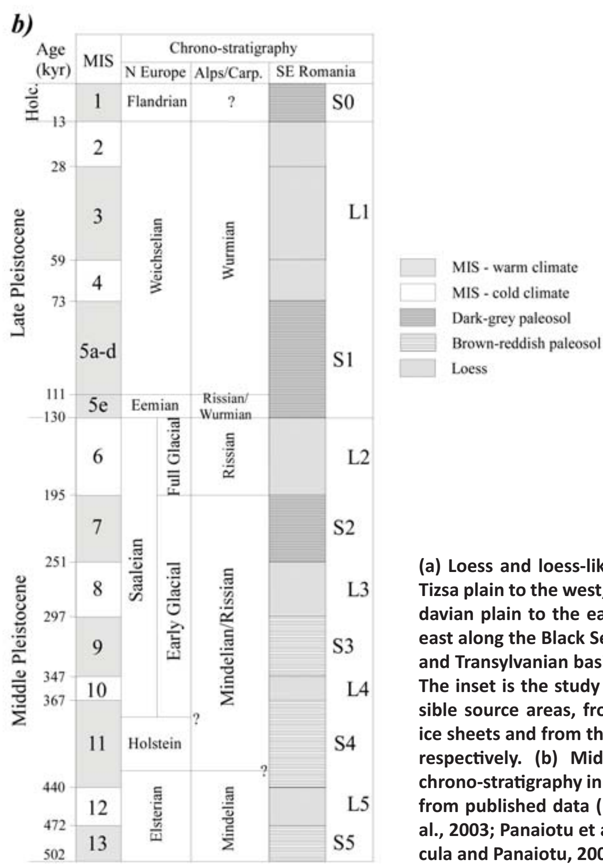
Elevations of the terrace erosional surfaces to the west in the Braşov intramontane basin (left box in Fig. 4.1b) are mainly taken from a study of Ghenea et al. (1971). This literature data is integrated with the new geomorphological observations performed during this study (Table 4.1b). Six terrace levels have been identified here (levels TI-TVI; Fig. 4.2b), with present-day elevations comprised between 100 and 10 m above the riverbeds (Fig. 4.3b), the older levels being strath-terraces, while the younger ones are fill-terraces (Table 4.1b). Denudation surfaces are present within the upper deposits of the fill-terraces.

The other factor important in the quantification of geomorphological evolution is the formation time of terrace bedrocks and denudation surfaces. Age data in this study comes from three different sources: new (N), derived (D) and traditional/previous (T) data (Table 4.1a-b). The new data (N; newly presented in this chapter) are obtained from dating the loess sequences covering the river terraces. The terrace bedrocks/denudation surfaces may or may not be directly overlain by the loess sequences. The sampling sites are placed at the base of the loess sequence. This study also gives newly derived age data (D) that have resulted from extrapolation the best available constraints of a river incision rate obtained at different levels of older terraces. Lastly, the traditional data (T) refers to previous published data in the literature, e.g. Grumăzescu (1973) for the Focşani basin and Ghenea et al. (1971) for the Braşov basin.

In the east, terrace ages have been derived as follows: level T1a by extrapolating the loess sequence dated in this study downstream in the Milcov valley combined with traditional data (Figs. 4.6a and 4.9; Table 4.1a), levels T1b-c to T8 by extrapolating the river

**Figure 4.4**




**(a) Loess and loess-like deposits in Romania occur in: Tizsa plain to the west, Danube plain to the south, Moldavian plain to the east, Dobrogea area to the south-east along the Black Sea coast and locally in the Braşov and Transylvanian basins (modified after Conea, 1969). The inset is the study area. Black arrows indicate possible source areas, from the north from the northern ice sheets and from the east from the Moldavian plain, respectively. (b) Middle Pleistocene-Holocene loess chrono-stratigraphy in the SE part of Romania compiled from published data (Panaiotu et al., 2001; Bălescu et al., 2003; Panaiotu et al., 2004; Necula et al., 2005; Necula and Panaiotu, 2008).**

incision rate obtained for level T1a (Fig. 4.9 and Table 4.1a), level T9 was dated during this study (Figs. 4.6c-d and 4.9; Table 4.1a), levels T10-11 by extrapolating the river incision rate obtained for level T9 (Fig. 4.9 and Table 4.1a). In the west, terrace ages are as follows: levels TI to TIII based on traditional chronology (Figs. 4.3b and 4.9; Table 4.1b), level TVIa-b dated during this study (Figs. 4.7a-b and 4.9; Table 4.1b) and levels TV-VI based on traditional chronology (Figs. 4.3b and 4.9; Table 4.1b). Further details regarding the way how the new and derived data were obtained and the error bars are given in Subchapter 4.6 and Table 4.3a-b.

### 4.3.2. Loess-paleosol sequences in Romania

Loess, defined as terrestrial clastic sediment, is predominantly composed of silt-sized particles and essentially an accumulation of wind-blown dust (Pye, 1995). In Romania, the loess and loess-like deposits cover mainly the low land plains: Tizsa plain in the western part, Danube plain in the south, Moldavian plain in the east and Dobrogea area in the south-east along the Black Sea coast. Loess deposits are also found locally in the Transylvanian and Braşov basins (Fig. 4.4a; Conea, 1969). Loess thicknesses are variable in these regions, reaching up to 50 m in the south-eastern part of the Danube plain and on the eastern flank of the Focşani basin (Dumitrescu et al., 1970), but decreasing down to 15-20 m along the Black Sea shore and even less towards the mountain chain.

Several previous studies have tried to establish a loess chronology and stratigraphy. These are firstly based on geomorphological analyses of the Quaternary terraces and the number of the loess-paleosol horizons accumulated on these terraces (Conea, 1969). More recently, a correlation with Marine Isotope Stages (MIS) has been suggested based on the magnetic measurements, IRSL analyses, magnetic proprieties and dynamic simulations carried out in the Mostiştea locality profile (SE of the Danube plain; Panaiotu et al., 2001; Panaiotu et al., 2004; Necula et al., 2005; Necula and Panaiotu, 2008) and on the IRSL analyses performed in the Tuzla locality profile (Black Sea shore; Bălescu et al. 2003; Fig. 4.4a).

Compiling the above data, several paleosol layers are interbedded within the loess deposits, from which only five will be used during this study because they fit to the Middle Pleistocene-Holocene period, which was studied. The two first oldest paleosols (S1-S2) are dark-grey (chernozemic type) corresponding to MIS5 and MIS7, while the next three (S3 to S5) are brown-reddish (brown forest type), being attributed to MIS9 to MIS13 (Fig. 4.4b). It is known that deposition of loess-paleosol sequences takes place under specific conditions. Loess accumulates preferentially during the cold climate (glacial stage), whereas paleosols form during a warmer period (interglacial stage; Starkel, 2003; Pan et al., 2003). In the study area, the Middle Pleistocene-Holocene loess can be sourced from two different areas, e.g. from the north from the northern ice sheets and/or from east from the Moldavian plain (red arrows in Fig. 4.4b). Local additional material might have been also supplied by the exhuming orogenic nappes. In the eastern part of the study area, loess thickness along the Putna valley increases from the orogen towards the foreland basin (Fig. 4.2a; Necea et al., 2005).

## 4.4. Processes controlling fluvial incision, terrace formation and loess deposition

The above address on terrace erosional surfaces and loess deposits target to determine river incision rates. However, important for the purpose of this research is to recognize, and where possible distinguish, between controlling factors, such as tectonics, climate and sea level change. In this regard, some relevant river network features are discussed for the SE Carpathians during the Middle Pleistocene-Holocene, which might have been controlled by such factors.

### 4.4.1. River analysis as an indicator of tectonic activity

The river network mirrors adequately the evolution of a mountain chain and this is reflected by the channel morphology and shape and geomorphological features (e.g. river terraces) developed along the river banks (Schumm et al., 2000). Evolution of a river channel is directly influenced by the vertical differential movements taking place in the orogen and adjacent basins. The relationship between the channel shape and the underlying bedrock in the SE Carpathian rivers has been previously presented in Chapter 2 and discussed in more detail in several studies (Ghenea et al., 1971; Posea, 1981; Rădoane et al., 2003; Fielitz and Seghedi, 2005).

In summary, a river displays a concave shape when it reaches the equilibrium condition (incision equals uplift), while the convex shape occurs in non-equilibrium (incision is succeed by the uplift; Burbank and Anderson, 2000; Schumm et al., 2000). In the study area, the present-day drainage divide separates the rivers in two groups, an eastern one with rivers flowing eastward through the Focşani basin and a western group of rivers that drains the Braşov intramontane basin (Fig. 4.5a; river parameters in Table 4.2).

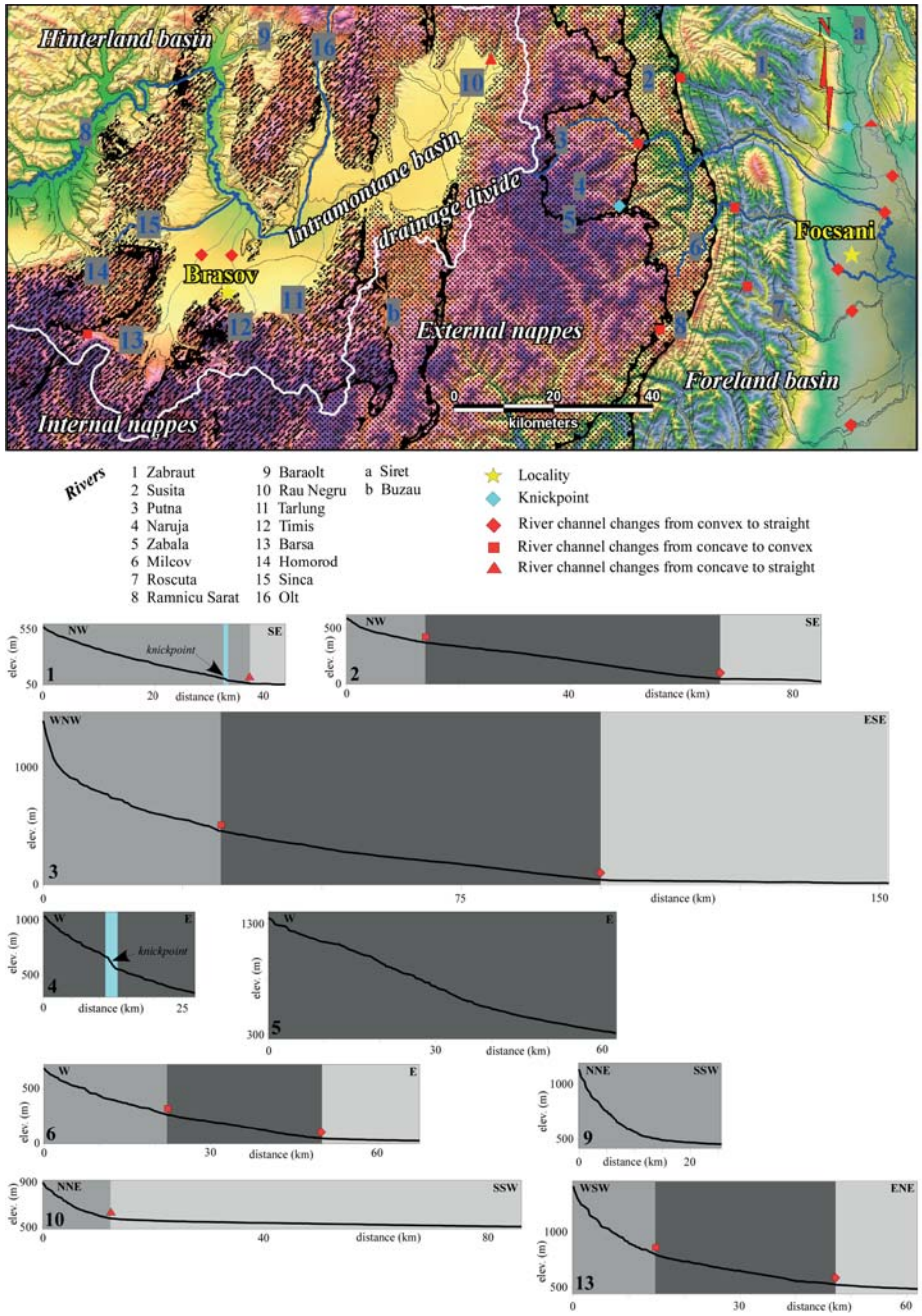
The river headwaters have a concave shape in the area to the east that contain the most-external nappes and the Upper Miocene-Lower Pliocene deposits of the foreland basin. This possibly indicates that the rivers are presently in equilibrium (up to red squares, Fig. 4.5). The middle reaches of the river profiles display a convex shape (Fig. 4.5a) and point to a non-equilibrium condition for the river sections through the Upper Pliocene-

**Table 4.2**

| Nr. | River name | Flowing direction | Channel length (km) | Channel shape |
|---|---|---|---|---|
| 1 | Zabraut | NW-SE | 43 | concave/straight |
| 2 | Susita | NW-SE | 85 | concave/convex/straight |
| 3 | Putna | W-E to NW-SE | 150 | concave/convex/straight |
| 4 | Naruja | W-E | 27 | convex |
| 5 | Zabala | W-E | 63 | convex |
| 6 | Milcov | W-E | 67 | concave/convex/straight |
| 7 | Roscuta | W-E | 50 | concave/convex/straight |
| 8 | Ramnicu Sarat | W-E | 121 | concave/convex/straight |
| 9 | Baraolt | NNE-SSW | 25 | concave |
| 10 | Rau Negru | NNE-SSW | 85 | concave/straight |
| 11 | Tarlung | S-N | 37 | concave |
| 12 | Timis | S-N | 33 | concave/straight |
| 13 | Barsa | WSW-ENE | 61 | concave/convex/straight |
| 14 | Homorod | W-E | 28 | convex |
| 15 | Sinca | SE-NW | 36 | concave |
| 16 | Olt | N-S to E-W | 195 | concave/straight |

**Names, flowing directions, channel lengths and present-day channel shapes of rivers draining the SE Carpathians. In the east, rivers have convex to concave shapes (numbers 1-8) draining eastward the Focşani foreland basin, while in the west rivers have concave channels (numbers 9-16) flowing westward through the Braşov intramontane basin (interpretation in text).**

Figure 4.5




**Main rivers analyzed for channel shape and morphology and their relationship with the bedrock. Numbers 1-8 are rivers draining eastward the external orogenic units and the Foşani foreland basin, while numbers 9-16 are rivers crossing the internal units and the Braşov intramontane basin. Notice the convex to concave channels in the east (profiles 1 to 6) and dominant concave channels in the west (profiles 9, 10 and 13), respectively. Rivers "a-b" are not used for interpretation. The present-day drainage divide is represented in white colour.**

Upper Holocene deposits. The river channels become straighter downstream in the Upper Holocene deposits (downstream from red diamonds; Fig. 4.5), corresponding to the area of present-day maximum subsidence. In the west, the rivers display dominantly the concave shape in the headwaters, becoming straight downstream at the points where the rivers enter in the intramontane basin (red diamonds, Fig. 4.5a). The straight shape is maintained until the confluence with the Olt main drainage river, area which corresponds to the most recent subsidence.

As the river gets older should increase its concavity in the upstream areas, approaching the equilibrium profile which corresponds to stable tectonic areas, while a younger river in active tectonic areas will display a convex, non-equilibrated shape (Rădoane et al., 2003). The former have been recognized along the Bistrița river (located far north of the study area), Ialomița and Dâmbovița rivers to the south (Rădoane et al., 2003) and Baraolt, Râu Negru and Bârsa rivers in the west in intramontane basin (Fig. 4.5b). The convex shapes are present, from north to south, along the Trotuş river (just north of the study area; Rădoane et al., 2003), Zăbrăuț, Şuşița, Putna, Zăbala, Năruja and Milcov rivers (Fig. 4.5b).

Two knickpoints have been observed along the Zăbrăuț and Năruja rivers (blue diamonds in Fig. 4.5a and grey shaded zone in Fig. 4.5b), which do not correspond to river confluence or change in lithology. They might correspond to a normal fault plane and to the frontal thrust of the Marginal Folds nappe (possibly reactivated?) due to recent tectonic activity.

River terraces studied in the adjacent sub-parallel valleys of the external SE Carpathians and foreland basin display distinct terrace geometries and configurations. The Putna valley shows a stair of well-developed terraces, with the highest and oldest level located in the orogenic nappes. Instead, the Şuşița valley located 2-3 km further northward has only poorly preserved terraces and no terraces are present along the Zăbrăuț valley further to the north (Fig. 4.2a). In the Milcov valley (south of Putna valley), terrace levels are locally preserved in the foreland basin and slightly better in the external nappes. One may consider that the largest river incision was reached by the Putna river when cutting through the orogenic nappes and decreasing progressively towards the foreland basin.

### 4.4.2. Climate and sea level changes

*Climatically*-driven river incision is generally based on the changes in slope, discharge, sediment load and calibre, erodibility of the bedrock (Gibbard and Lewin, 2009). The sediment/discharge ratio occurs when low sediment load is coeval with high water discharge. Gibbard and Lewin (2009) suggest that enhanced fluvial activity is a result of the occurrence of cold (glacial) climates and the supply of abundant fresh materials by periglacial slope processes, intensified during the 'mid-Pleistocene transition' (1.2-0.8 Ma). Periglacial period refers to transition from cold to warm climate and vice versa. The glacial period corresponds also to accumulation of loess deposits. During the interglacial period (warmer climate), the river cannot erode its channel bed because it is not supplied with enough water discharge to move the coarse material due to vegetation growth on hillslope (Maddy et al., 2000; Pan et al., 2003; Gibbard and Lewin, 2009), which results in formation of paleosols.

Terrace erosional surfaces that formed during the glacial and periglacial periods might be controlled by climate, whereas tectonic control appears more evident for those terraces that formed during interglacial climate (i.e. steady warmer). Strath-terraces develop only on one side of the river channel, known as unpaired terraces and are formed due to the lateral erosion of the river into the bedrock associated with uplift and tilting (Burbank and Anderson, 2001). The river incision might be enhanced during the climatic transitions due to increase in water discharge. However due to their evolution, these terraces may be covered only by a thin layer of coarse channel-bed sediments. Fill-terraces are well developed on both river banks and occur as paired terraces (Burbank and Anderson, 2001). Fill-terraces form by multiple vertical incisions of the river channel into the valley filling during cold-warm transition when the water discharge increases due to regional climatic events (Merritts et al., 1994; Pan et al., 2003). Furthermore, cold-warm transition might be correlated with flooding events, when thickness of the sedimentary layer at the bottom of the riverbed corresponds to the depths of the river pools. In the case of the studied Putna, Milcov, Homorod and Olt rivers, pools depth are estimated to be in the range of few meters.

*Sea-level* drops combined with climate events are known to enhance erosion in source areas, such as the shift in exhumation in the orogenic core of the Alps during the Messinian Salinity Crisis (Willett et al., 2006). A coeval and comparable event took place in the Black Sea around 6-5 Ma, which led to the massive desiccation of the Carpathians foreland (Gillet et al., 2007) and potential erosion in the Carpathian core as shown by the low-temperature thermochronology (Merten et al., accepted).

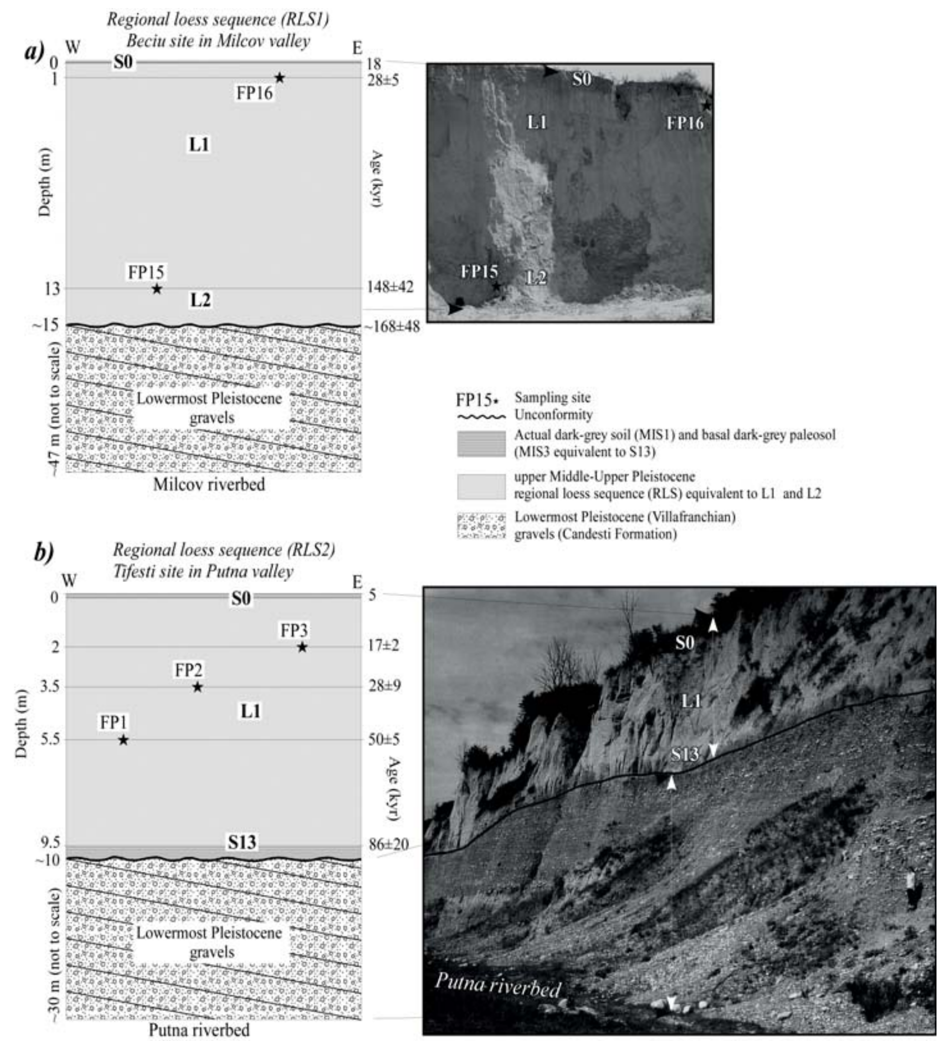
Furthermore, cold and cold-warm transition might result in sea-level fluctuations that with base-level change may also have an affect on river-incision. Glacial interval corresponds to regression phase (decrease of sea level), while interglacial to transgression due to melting of glaciers (increase of sea level). During the glacial interval, the sediment supply might increase and therefore the material would be removed from the land areas to be deposited in marine basins (Gibbard and Lewin, 2009). The sea-level drop could be reflected in river incision accompanied by increased denudation of emergent surfaces.During the Quaternary, the most significant fluctuation of the Black Sea level took place during the last glaciation, early Middle Pleistocene (MIS16), when the sea level decreased with 150 m (Winguth et al., 2000). This indicates a dry climate in the Black Sea region, when evaporation was probably higher than precipitation and fluvial water influx. In the Early Ceaudian (Gunz-Mindelian interglacial), the sea level raised with 100-130 m relative to the present-day sea level. Both the sea level drop and subsequent rise took place during the early Middle Pleistocene (~592-627 kyr), somehow coeval with formation of the oldest river terrace in the SE Carpathian external nappes as reported by Grumăzescu (1973). The river terrace levels in the study area are located between 650 and ~200 m relative to the present-day sea level (Fig. 4.3). The sea-level fluctuations of the Black Sea during the Middle Pleistocene-Holocene might be excluded as a steering parameter of enhanced river incision and terrace formation (Winguth et al. 2000; Aksu et al. 2002 and references therein) because the Black Sea basin was located ~170 km to the east of the study area and neither of the studied rivers were discharging directly into the Black Sea basin.

## 4.5. Sampling scheme and age determination

### 4.5.1. Sample locations

This study provides ages for loess deposits and indirectly for formation of terrace levels or denudation surfaces. These ages come from the optical dating applied to loess-paleosol sequences, which cover both abandoned strath- and fill-terraces and tilted Low-

**Figure 4.6a-b**




**Cross-sections show regional loess sequences in the Milcov and Putna valleys, Focşani foreland basin (Beciu and Ţifeşti sites; locations in Fig. 4.2a). (a) Beciu site yielded IRSL ages of 28±5 kyr (1 m) and 148±42 kyr (13 m), a loess accumulation rate of 0.1±0.03 mm/yr (Fig. 4.8), a derived age for basal loess of 168±48 kyr (minimum age for denudation surface placed at ~47 m) and a computed river incision rate of 0.3±0.08 mm/yr (Table 4.3). (b) Ţifeşti site yielded IRSL ages of 17±2 kyr (2 m), 28±9 kyr (3.5 m) and 50±5 kyr (5.5 m), a mean loess accumulation rate of 0.1±0.04 mm/yr, a derived age for basal loess/paleosol S13 of 86±20 (minimum age for denudation surface placed at ~30 m and a computed river incision rate of 0.3±0.08 (Table 4.3a). The 500 m distant sample locations on this continuous outcrop, have been projected to this site to better show their relation to the underlying sedimentary succession.**

ermost Pleistocene strata in the foreland basin. Not the terrace levels themselves, nor their coarse-grained covers can be directly dated. Instead, deposition ages of loess-paleosol sequences have been determined to provide constraints on river incision history. The applied dating technique is infrared stimulated luminescence (IRSL), which determines the last exposure of the mineral grains to daylight prior to deposition (see Chapter 3 for details). It requires that the grains experienced sufficient bleaching during transportation and sedimentation in order to reset the luminescence signal to a zero value. This prerequisite was fulfilled for the wind-blown sediments (i.e. loess deposits in this study), but not for water-laid fluvial sediments, due to the poor light conditions during formation. This study initially sampled both loess deposits and fine-grained fluvial sediments, the former gave good IRSL results, while the latter not and have been omitted in the rest of this study.

The sampling strategy was to take as many samples as possible from as many terraces as possible, starting with the base of the loess sequences. In the thicker loess deposits, additional samples have been taken to evaluate the age of the complete sequence. 22 samples have been collected in total along the W-E profile (A-A' profile in Fig. 4.1b), from which 11 samples gave reliable results and have been further used for interpretation. In the east, the sampling was concentrated along the Putna and Milcov rivers (Fig. 4.2a) due to the well-preserved and well-accessible loess deposits (e.g. regional sequences, loess-covered terraces). The sampling of this area was also important because it marks a larger amplitude of the uplift during the late Early Pleistocene (~750 m) and Middle Pleistocene-Holocene (250 m). Fewer samples have been taken in the west in the Braşov intramontane basin due to less suitable sediments, the terraces being more often covered by fluvial sands (Fig. 4.2b). All samples have been processed according to the standard sample separation and analysis protocol used at the Heidelberg luminescence laboratory, which is based on Lang et al. (1998) and Kadereit (2000; see details in Chapter 3).

### 4.5.2. Determination of loess deposition rates, terrace ages and river incision rates

The IRSL ages obtained in this study mainly correspond to the base of the loess-paleosol sequences and reflect the time when the loess accumulated either over the terrace bedrocks or other denudation surfaces. These ages have been compared with Marine Isotope Stages (MIS), chrono-stratigraphy of North Europe and Alps/Carpathians systems and loess chrono-stratigraphy of the SE Romanian area (Fig. 4.8).

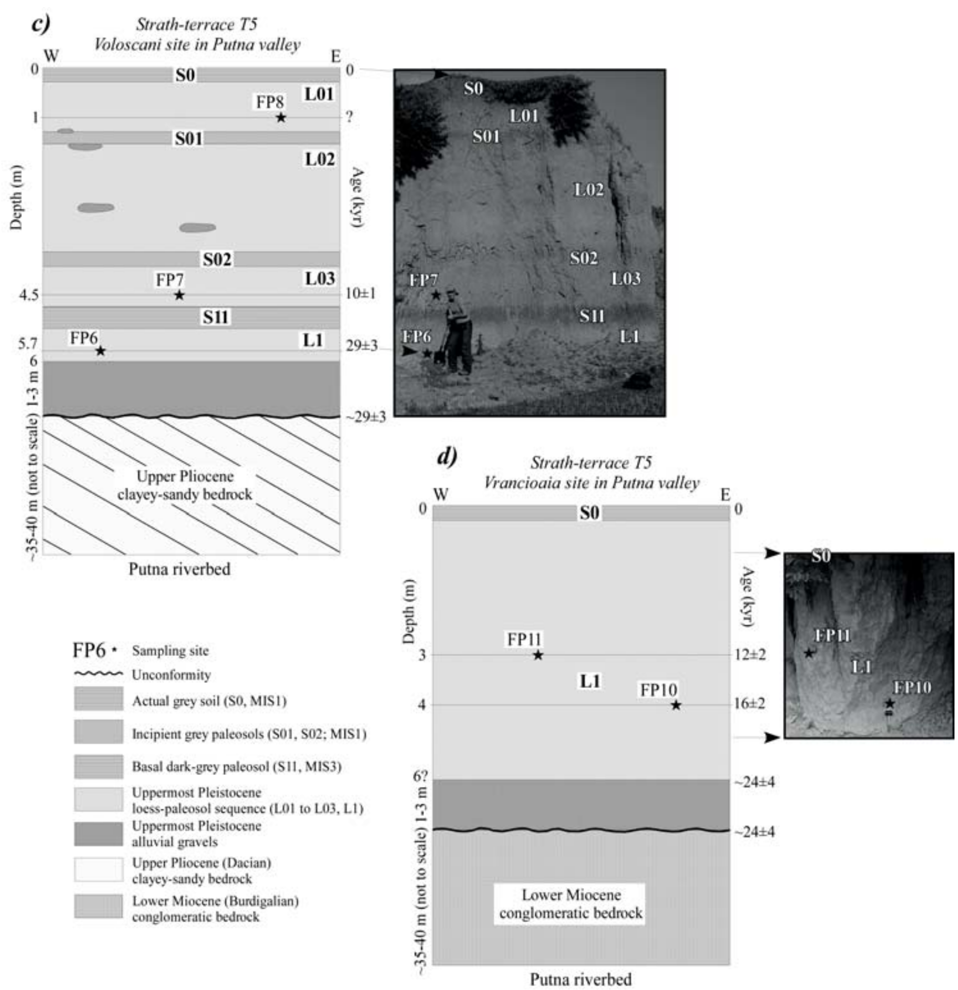
An undetermined time period might exists between the deposition of the youngest fluvial sediments and the oldest aeolian accumulations (e.g. level T9), thus the latter can be deposited only after the river left the former fluvial level (riverbed) and started to incise at a deeper level, leaving behind an abandoned terrace level. Assuming that the loess accumulated over the fluvial deposits immediately or shortly after the river abandoned the terrace level, then the extrapolated age gives a minimum formation age for the terrace bedrock or denudation surface, respectively (Fig. 4.8 and Table 4.3a-b).

The accumulation rates have been extrapolated down to the base of the loess sequence in order to obtain a minimum age for loess deposition (Fig. 4.8; Pan et al., 2003). The error bar for the accumulation rate represents a certain percentage from the rate itself, this percentage is equal to that percentage of the error bar of the basal loess age. Error bar for the river incision rate is a certain percentage from the rate itself, which is equal to that percentage of the error bar of the terrace age.

Loess accumulation rates have been calculated either between two IRSL ages (Beciu, Țifeşti and Vrâncioaia sites; Figs. 4.6a-b, d and 4.8) or between one age and the upper limit of the glacial/interglacial stage (MIS) to which it corresponds (Ariuşd and Arini sites; Figs. 4.7a-b and 4.8). Lastly, in the case of the Voloşcani site (Fig. 4.6c), an accumulation rate was not calculated between the two IRSL ages because of the paleosol horizon present in between them, which might introduce a very high error in the accumulation rate.

Few factors have been considered when the ages were calculated and interpreted: the accumulation rate can change in time (sites with 2 or more measurements), long- or short-term fluvial reworking and redeposition shown by the intercalated coarser lenses

**Figure 4.6c-d**

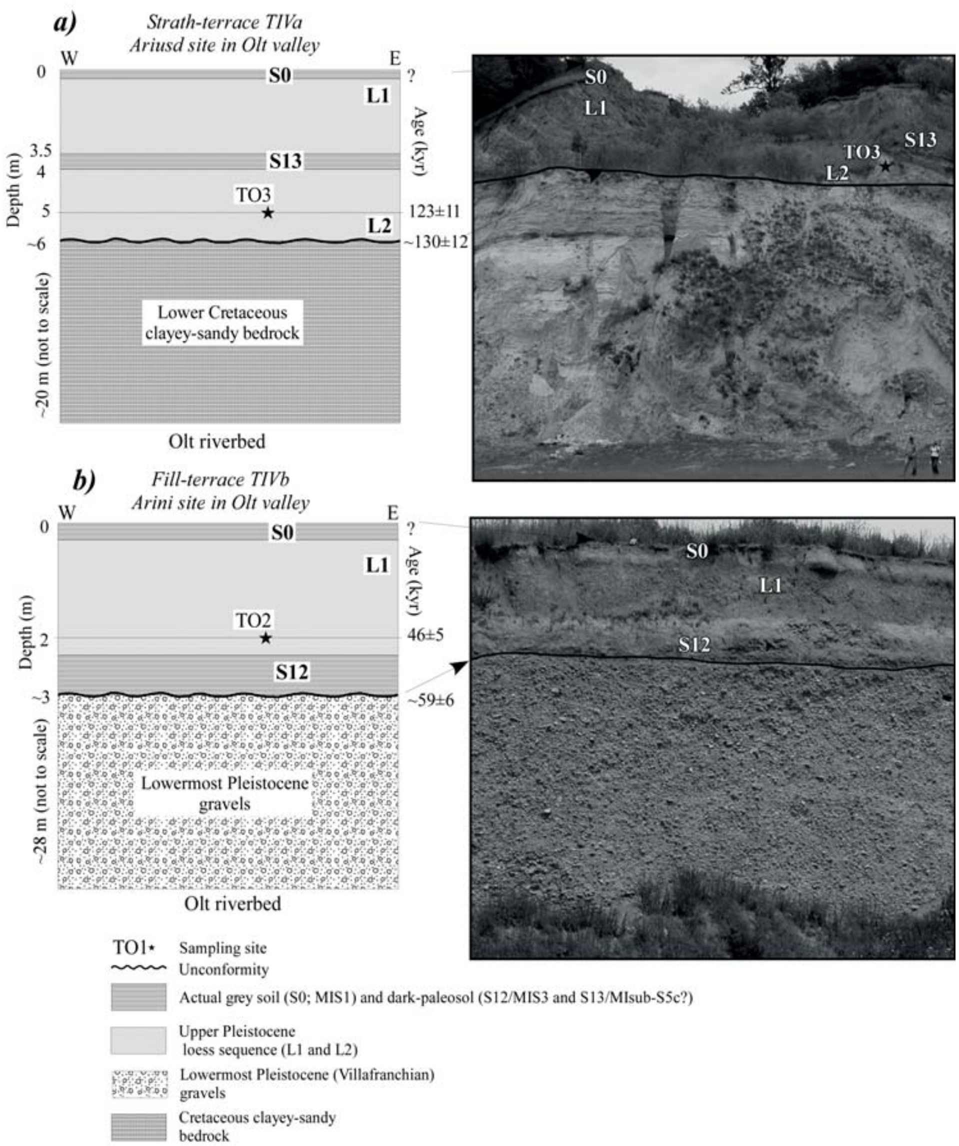



**Cross-sections show strath-terrace T5 (39-41 m) in the Putna valley, Focşani foreland basin (Voloşcani and Vrâncioaia sites; locations in Fig. 4.2a). (c) Voloşcani site yielded IRSL ages of 10±1 kyr (4.5 m) and 29±3 kyr (5.7 m), the latter is minimum formation time for terrace bedrock placed at 35-40 m. The river incision rate is 1.3±0.13 mm/yr (Table 4.3a). (d) Vrâncioaia site yielded IRSL ages of 12±2 kyr (3 m) and 16±2 kyr (4 m), a loess accumulation rate of 0.25±0.005 mm/yr (Fig. 4.8), a derived age for basal loess of 24±4 kyr (minimum age for terrace bedrock) and a computed river incision rate of 1.5±0.2 mm/yr (Table 4.3a).**

(e.g. the upper part of the loess sequence covering level T9), presence of paleosol horizons within the loess sequences (e.g. level T9), periods of non-deposition represented by the erosional surface(s) within the loess sequences (is not the case), possible longer time-span between formation of terrace bedrock and onset of loess deposition.

Loess stratigraphy cannot be accurately evaluated because of the restricted number

**Figure 4.7**




**Cross-sections show strath-terrace TIVb (26 m) and fill-terrace TIVa (30-40 m) in the Olt valley, Braşov intramontane basin (Ariuşd and Arini sites; locations in Fig. 4.2b). (a) Ariuşd site yielded an IRSL age of 123±11 kyr (5 m), a derived age for basal loess of 130±12 kyr (minimum age for terrace bedrock placed at 20 m) and a computed river incision rate of 0.15±0.01 mm/yr (Table 4.3b). (b) Arini site shows an IRSL age of 46±5 kyr (2 m), a basal paleosol S12 of <59±6 kyr (minimum age for denudation terrace surface placed at 28 m) and a computed incision rate of 0.5±0.03 mm/yr (Table 4.3b).**

of the sampling points and also due to the lack of the additional data. Neither detailed stratigraphic nor sedimentological research has been performed during this study because its main focus is to derive the processes controlling terrace formation and river incision, not to establish the chrono-stratigraphy of the loess and fluvial deposits. The accumulation rates calculated between different levels in the loess deposits were extrapolated to the deeper parts of the outcrops. In the case of terrace bedrocks covered by fluvial deposits, the extrapolation was done down to the top of the fluvial sediments (Voloşcani and Vrâncioia sites; Fig. 4.6c-d), while for those surfaces covered directly by loess, the extrapolation was done until the bedrock surfaces (Ariuşd site; Fig. 4.7a). In the case of denudation surfaces, the bases of the dated loess sequences are in direct contact with theses surfaces (Țifeşti and Beciu sites in Fig. 4.6a-b; Arini site in Fig. 4.7b). In the Voloşcani site, the loess deposition might be discontinuous due to the presence of the intercalated paleosol horizons, but in the Ariuşd, Țifeşti, Beciu and Arini sites the loess sedimentation is quasi continuous which allows to consider that the IRSL minimum ages for loess deposition in these sites are accurate enough to be used to date the river terraces.

After determining the minimum formation age for terrace levels and denudation sur-

**Table 4.3**

***a)***

| Terrace level and regional loess seq./ RLS (occurrence) | Loess accuml. rate (mm/yr) | Terrace age (kyr) | River incision/ uplift rate (mm/yr) | Corresponding MIS and glacial (this study) | Climate |
|---|---|---|---|---|---|
| RLS1 (foreland) | 0.1±0.03' | 168±48 | 0.3±0.08* | MIS6 (Rissian) | Cold |
| T1a (orogen) | - | 168±48 | 1.4±0.4* | MIS6 (Rissian) | Cold |
| T1b (orogen) | - | 141±39 | - | MIS6 (Rissian) | Cold |
| T1c (orogen) | - | 108±30 | - | MIsub-S5d (Würmian) | Warm? |
| RLS2 (foreland) | 0.1±0.04' | 86±20 | 0.3±0.08* | MIsub-S5c? | Cold? |
| T2 (orogen) | - | 72±20 | - | MIS4 | Cold |
| T3 (orogen+foreland) | - | 43±12 | - | MIS3 | Warm |
| T4 (orogen+foreland) | - | 35±10 | - | MIS3 | Warm |
| T5 (foreland) | - | 29±3 | 1.3±0.13* | MIS3/MIS2 | Warm/Cold |
| (orogen) | 0.25±0.005' | 24±4 | 1.5±0.2* | MIS2 | Cold |
| T6 (orogen+foreland) | - | 16±2 | - | MIS2 | Cold |
| T7 (orogen+foreland) | - | 11±1 | - | MIS2/MIS1 | Cold/Warm |
| T8 (orogen+foreland) | - | 5±0.5 | - | MIS1 | Warm |
| T9 (orogen+foreland) | - | 2±0.2 | - | MIS1 | Warm |
| T10 (orogen+foreland) | - | <2±0.2 | - | MIS1 | Warm |
| T11 (orogen+foreland) | - | <2±0.2 | - | MIS1 | Warm |

***b)***

| Terrace level and regional loess seq./ RLS (occurrence) | Loess accuml. rate (mm/yr) | Terrace age (age of MIS) (kyr) | River incision/ uplift rate (mm/yr) | Corresponding MIS or glacial (this study) | Climate |
|---|---|---|---|---|---|
| TI (orogen) | - | 487±15 (472-502) | 0.2±0.006^ | MIS13 (Mindelian I) | Warm |
| TII (orogen) | - | 456±16 (440-472) | 0.15±0.005^ | MIS12 (Mindelian II) | Cold |
| TIII (intramontane) | - | 162±32 (130-195) | 0.3±0.06^ | MIS6 (Rissian I) | Cold |
| TIVa (orogen) | - | 130±12 | 0.15±0.01* | MIS6/MIsub-S5e | Cold/Warm |
| TIVb (intramontane) | - | 59±6 | 0.5±0.03* | MIS4/MIS3 | Cold/Warm |
| TV (intramontane) | - | 92±19 (73-111) | 0.2±0.04^ | MIsub-S5a-d? (Würmian I) | Warm-Cold |
| TVI (intramontane) | - | 20±7.5 (13-28) | 0.3±0.1^ | MIS2 (Würmian III) | Cold |

**Loess accumulation rate, terrace age, river incision/uplift rate and corresponding MIS (and glacial stage where applicable) and type of climate when each terrace was formed (a) at the transition zone from the Carpathian orogen to the Focşani basin and (b) in the Braşov intramontane basin. 0.1±0.03' is the accumulation rate with error bar. 0.3±0.08* and 0.2±0.006^ are the river incision rates with error bars derived by using the new IRSL data from this study and those ages published by Ghenea et al. 1971, respectively.**

faces, these data may be used to derive the river incision rates. This study addresses the maximum incision rates because the loess might not have accumulated immediately after the river has abandoned the former terrace level. The vertical river incision might be directly proportional to the uplift rate in an active orogen (Merritts et al., 1994). The river incision/uplift rate was extracted using the linear relationship between the elevation of each bedrock surface and its minimum age (Burbank and Anderson, 2001):

$$I = h / t$$

where, $I$ is the rate of river incision into the bedrock, $h$ is the relative vertical distance calculated between terrace bedrock surface and actual riverbed and $t$ is the minimum computed age of the terrace bedrock surface.

The maximum river incision rates plotted along longitudinal river profiles and geological cross-section will help to determine the processes controlling the river incision into the bedrock, which might be a coupling between tectonic, climatic and erosional processes (Whipple and Tucker, 1999, 2002; Whittaker et al., 2007).

## 4.6. Results and interpretations

### 4.6.1. External Carpathian orogen and foreland basin

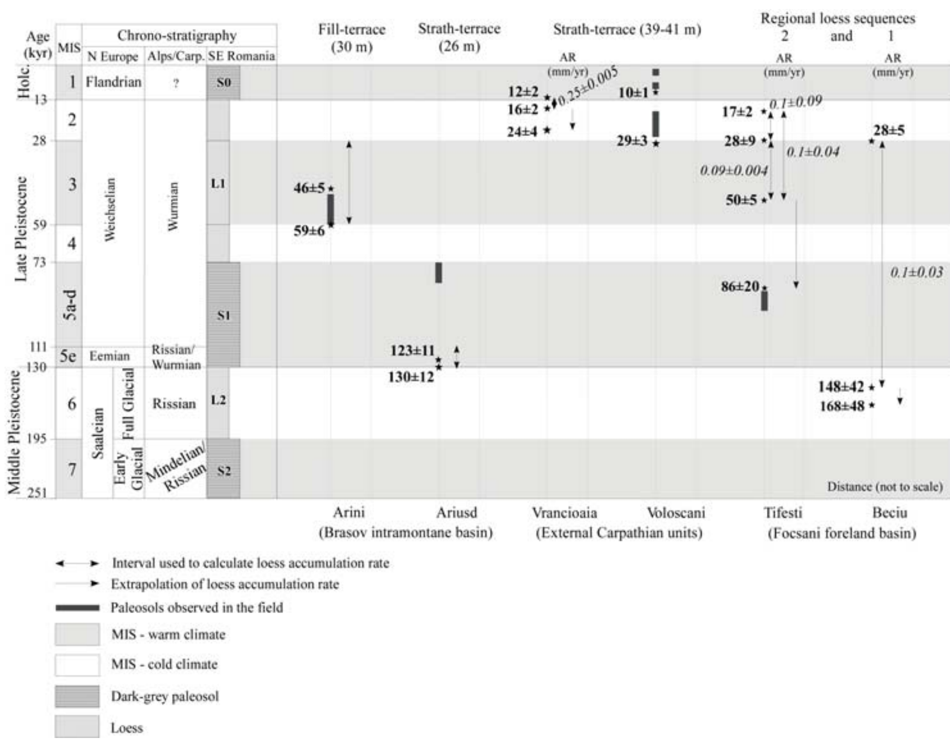
The regional loess sequences (RLS) present on the western flank of the Focşani basin have been dated during this study and analysed together with the eleven terrace levels previously mapped in the transition zone from the Carpathian orogen to the Focşani basin. The river terraces are located at elevations comprised between 250 and 1 m above the riverbed, being mostly strath- and partly fill-terraces (Fig. 4.2a; Table 4.1a).

#### *4.6.1.1. Middle Pleistocene: upper strath-terrace and regional loess sequence (RLS)*

The Middle Pleistocene sediments are represented by the gravely deposits of the strath-terrace level T1 and by the overlying aeolian sediments, the regional loess sequence (RLS). This terrace is composed of sublevels T1a (247 m), T1b (190-220 m) and T1c (159 m; Table 4.1a), which are levelled into the Miocene molasses sediments of the Subcarpathian nappe, forming the Bârseşti plateau (Fig. 4.2a). Their sedimentary covers are generally composed of 3-7 m basal gravels and fine clay (Table 4.1a), locally covered by loess-like deposits. Based on the traditional terrace chronology, Dumitrescu et al. (1970) attributed level T1a to the Middle Pleistocene, partially being covered by a loess-like sequence of 12-15 m. The same study correlates this sequence with a sequence extending further eastward in the foreland basin along the Milcov valley, based on the reddish paleosol layers observable on the left bank of this river.

During this study, the loess sequence covering level T1a has not been observed in the field. The samples for IRSL dating were collected from the well-exposed section on the right bank of the Milcov river, in the Beciu site (samples FP15-16; Figs. 4.3a and 4.6a), where the RLS covers unconformably the Lowermost Pleistocene lacustrine conglomerates. Samples collected at 1 and 13 m in depth yielded ages of 28±5 and 148±42 kyr, which correspond to MIS3/MIS2 transition (warm/cold climate) and MIS6 (Rissian glacial; Fig. 4.8). Neither recognizable paleosols nor erosional surfaces have been observed within this sequence, which allowed to compute an accumulation rate of 0.1±0.03 mm/yr. This rate was extrapolated to the base of the loess sequence, for which an age of ~168±48 kyr was obtained (Fig. 4.6a) that corresponds to the Rissian glacial stage (Fig. 4.8).

Figure 4.8



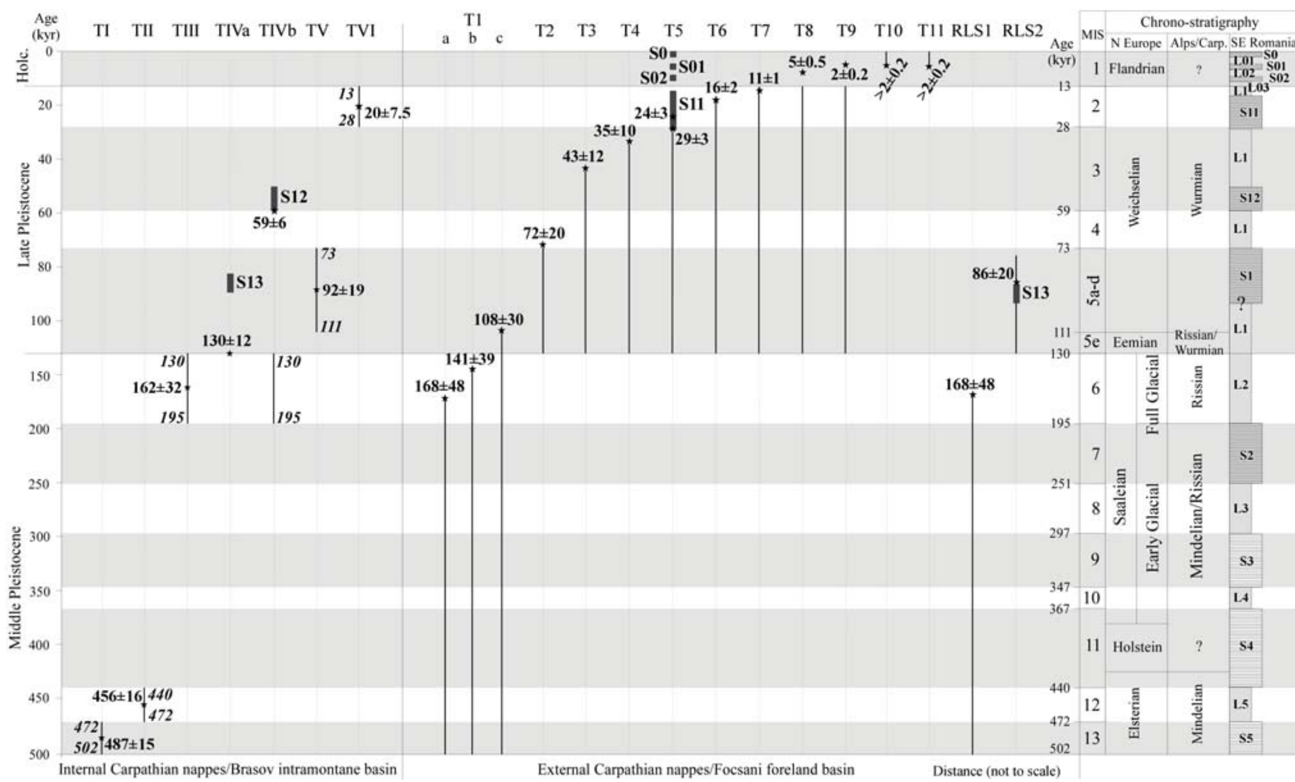

**New IRSL and extrapolated ages obtained in the SE Carpathians (sites from Figs. 4.6 and 4.7). Ages are correlated with Marine Isotope Stages (MIS) and chrono-stratigraphy of North Europe and Alps/Carpathians system (Late Pleistocene, Mol et al., 2000; Middle Pleistocene, Shackleton and Opdyke, 1973; Kukla, 2005) and loess chrono-stratigraphy for the SE part of Romania (Panaiotu et al., 2001; Bălescu et al., 2003; Panaiotu et al., 2004; Necula et al., 2005; Necula and Panaiotu, 2008). AR is the calculated accumulation rate (mm/yr) with error bar extrapolated down to the base of the loess sequences for each studied site. Grey thick vertical lines are paleosols observed in the field. Black bold and italic numbers are the IRSL ages and accumulation rates, respectively. Notice the change in the time scale at 100 kyr, for a detailed display of the stratigraphic column.**

In terms of loess chrono-stratigraphy, the entire sequence corresponds to L1 and L2 (Fig. 4.9) without observable paleosols or erosional surfaces in between them (Fig. 4.6a). The obtained results are younger than those previously published for this area (Middle Pleistocene, Saulea et al., 1968; Dumitrescu et al., 1970). Previous paleomagnetic dating indicates an age of ~400 kyr eastward in the Mostiştea site in the Danube plain (Fig. 4.4a; Panaiotu et al., 2001), while IRSL analysis performed by Bălescu et al. (2003) in the Tuzla site on the Black Sea coast suggests an age of ~300 kyr. Although, both sites are too far away from the study area to make reliable correlations.

Additional loess sampling was performed in the Ţifeşti site, on the left bank of the Putna river (FP1 to FP3; Figs. 4.3a and 4.6b). The computed ages are 17±2 kyr (2 m), 28±9 kyr (3.5 m) and 50±5 kyr (5.5 m), corresponding to MIS2 and 3 (cold and warm climates). The computed accumulation rates are 0.1±0.09 and 0.09±0.004 mm/yr, from which a mean rate of ~0.1±0.04 mm/yr has been obtained (Fig. 4.8). Extrapolating this rate down to the deeper part of the sequence suggests that the loess accumulated around 86±20 kyr during a short cold period (MIsub-S5c). The loess deposition occurred after formation of the basal grey paleosol (S13) during a warmer period (Figs. 4.6b and 4.9).

The denudation surfaces observed at the contact between the RLS and the Lower-

Figure 4.9




Formation times for terrace bedrocks and loess sequences in the SE Carpathians. Ages were correlated with MIS and chrono-stratigraphy of the North Europe-Alps/Carpathians systems to assign the corresponding climate, based on which the loess chrono-stratigraphy in Romania was revised (to compare see Fig. 4.4b). Black italic ages and black thin vertical lines represent the possible formation time for terraces (e.g. Middle Pleistocene for level T1) taken from published literature (references in Table 4.2a-b). Other black values are the new computed ages during this study from the orogen-foreland zone and intramontane basin (new, derived and traditional data in Table 4.3a-b). Grey thick vertical lines are paleosols identified in the field. Notice that terrace erosional surfaces were formed during either full climate (warm or cold) or at climatic transitions.

most Pleistocene deposits in the Milcov and Putna valleys are placed at ~47 and ~30 m above the river, having minimum ages of ~168±48 and 86±20 kyr, respectively (Fig. 4.6a-b). The maximum obtained river incision rates in the two sites are 0.3±0.08 mm/yr for the late Middle Pleistocene in the Micov valley and 0.3±0.08 mm/yr for the Late Pleistocene in the Putna valley, respectively (Table 4.3a).

Deposition time of the RLS was interpreted as the minimum time formation of terrace T1, thus, the bedrock surfaces of sublevels T1a, b, c could have minimum ages between 168±48 and 86±20 kyr. The bedrock surface of sublevel T1a (237-242/mean 239 m) has been tentatively correlated with the oldest age (168±48 kyr, Rissian glacial), revealing a maximum river incision rate of 1.4±0.4 mm/yr (Table 4.3a). Assuming that the river incision remained unchanged in time, then sublevels T1b (187-210/mean 198 m) and T1c (149-154/mean 151 m) may have formed around 141±39 and 108±30 kyr, which correlate with MIS6 (Rissian glacial) and MIsub-S5d (cold climate; Würmian glacial; Fig. 4.9 and Table 4.3a).

#### ***4.6.1.2. Late Pleistocene: middle strath-terraces***

The river adjustment to the actively uplifting orogen continued during the Late Pleistocene, reaching its maximum amplitude in the Subcarpathian nappe (strath-terraces T2-T5; Fig. 4.2a), while loess accumulation covered a larger area extending westward in the Braşov intramontane basin and probably until the Transylvanian basin. The Middle strath-terraces are represented by levels T2 (89-130 m), T3 (70 m), T4 (51-60 m) and T5 (39-41 m; Fig. 4.3a) and have a presumed Late Pleistocene age (Dumitrescu et al., 1970). Most of the terraces overlay only the Subcarpathian nappe, except for level T5, which covers also the Upper Miocene-Pliocene strata of the Focşani basin (Fig. 4.2a). Their bedrock surfaces are placed at 87-116/mean 101 m, 61 m, 46-53/mean 49 m and 35-40/mean 37 m, respectively, and are covered by 2-6 m fluvial gravely deposits and an upper sandy sequence of 1-3 m thickness (Table 4.1a; Grumăzescu, 1973).

No direct IRSL data have been obtained for levels T2-T4, but common characteristics with level T1 (strath-terrace type, same location) suggest that the uplift might have kept at a constant rate of 1.4±0.4 mm/yr during the Late Pleistocene. Consequently, terrace ages for levels T2-T4 have been calculated by dividing the bedrock elevations over the before uplift rate. Thus, level T2 formed around 72±20 kyr (MIS4, cold climate), probably after the basal paleosol from the Putna valley has formed (Ţifeşti site; Fig. 4.9), while ages of 43±12 and 35±10 kyr computed for levels T3 and T4 fit entirely to MIS3 (warm climate; Fig. 4.9 and Table 4.3a).

Next terrace level T5 (39-41 m) overlays largely the Subcarpathian nappe and the sedimentary succession on the western flank of the Focşani basin (Fig. 4.2a). Its bedrock placed at 35-40/mean 37 m above the Putna riverbed is covered by gravely deposits of 1-3 m and an upper loess-paleosol sequence of 6 m (Table 4.1a). This sequence was dated by IRSL method in the Voloşcani and Vrâncioaia sites (samples FP6-7 and FP10-11, respectively; Fig. 4.2a). In the Voloşcani site (Fig. 4.6c), the loess-paleosol sequence observed in the field is made up of 0.5 m recent soil (S0), 1 m loess (L01), two light-grey incipient paleosols of 0.3 m each (S01 and S02) separated by 1.7 m loess (L02) and two additional loess horizons of 1 m each (L03 and L1) separated by a dark-grey paleosol of 0.5 m (S11). The sandy-gravely lenses from horizons L01 and L02 indicate reworked material that are probably sourced from the uphill regions, while the angular blocky structures from horizons L03 and L1 point to an aeolian transport (photo in Fig. 4.6c).

The dated horizons L03 and L1 at terrace level T5 yielded ages of 10±1 kyr (4.5 m) and 29±3 kyr (5.7 m), which fit to MIS2/MIS1 and MIS3/MIS2 transitions (Fig. 4.8). Horizons S01 and S02 are interpreted as incipient paleosols, which formed during MIS1 (warm climate). An accumulation rate could not be computed between the two IRSL ages because of the presence of the paleosol horizon S11 in between them. The presence of thispaleosol will introduce a high error in the accumulation rate sincethe onset of loess deposition and terrace formation time are being overestimated. To avoid this, the oldest age of 29±3 kyr was chosen as the onset of loess deposition, which corresponds to MIS3/MIS2 transition (warm/cold). It thus follows from the above that S11 mainly formed during a cold climate, the cause of this cannot be explained at this point. However, the minimum formation time of level T5 can be placed around 29±3 kyr, in the upper part of MIS3 (warm climate).

The additional dating of terrace level T5 carried out in the Vrâncioaia site (Fig. 4.6d) revealed comparable ages of 12±2 (3 m) and 16±2 kyr (4 m), respectively, which fit to MIS2 (cold climate). An accumulation rate of 0.25±0.005 mm/yr was obtained (Fig. 4.8). This rate was extrapolated for a depth of ~2 m, resulting an age of 24±4 kyr (MIS2, cold climate), slightly younger than that one obtained in the Voloşcani site. Because the entire loess thickness is not well-known at this site, the terrace age might be less accurate. Concluding, level T5 formed around 29±3 and 24±4 kyr, which corresponds generally to MIS3/MIS2 transition (warm/cold climate) and MIS2. This age corresponds to one of the major terrace formation episodes in this area, which clearly formed long after the loess started to accumulate around 168±48 kyr. Bedrock elevation of 35-40/mean 37 m and its age (29±3 and 24±4 kyr, respectively) gave slightly variable (maximum) river incision rates, such as 1.3±0.13 mm/yr in the Voloşcani site and 1.5±0.2 mm/yr in the Vrâncioaia site (orogen), respectively (foreland basin; Table 4.3a).

#### *4.6.1.3. Latest Pleistocene-Holocene: lower strath- and fill-terraces*

There is no direct data obtained for the lower strath-terrace levels T6 (28-31 m), T7 (19-21 m), T8 (9-11 m) and T9 (5 m; Figs. 4.3a and 4.4a). The bedrock surfaces situated at 22, 15, 7 and 2.5 m are covered by basal gravels of 1-2 m and sand of 0.5-2 m (Grumăzescu, 1973; Table 4.3a). The Putna river incision during the Latest Pleistocene reached rates of 1.5±0.2 mm/yr in the Subcarpathian nappe and 1.3±0.13 mm/yr on the western flank of the Focşani basin as described above for terrace T5. Levels T6-T9 cover equally the orogen and the foreland basin, and therefore both rates have to be taken into account when terrace ages are calculated by using a mean rate of 1.4±0.16 mm/yr. This implies that level T6 was formed at 16±2 kyr, T7 at 11±1 kyr, T8 at 5±0.5 kyr and T9 at 2±0.2 kyr, the first two levels corresponding to MIS2/MIS1 transition (cold/warm climate), while the last two formed during MIS1 (warm climate; Fig. 4.9). The lowermost levels T10 and T11, fill-terrace type, are made up entirely of alluvial sediments and have a presumed Upper Holocene age (Dumitrescu et al., 1970) or are younger than 2±0.2 kyr (age of level T9; Fig. 4.9). They might have formed subsequently to levels T8 and T9 during the Latest Holocene because they are situated in a lower position or represent the actual alluvial plain of the studied rivers.

### 4.6.2. Internal Carpathian orogen and intramontane basin

The elevation of the Quaternary sediments is also important for understanding of the geomorphological evolution of the internal Carpathian nappes and Braşov intramontane basin. Lower Pleistocene sediments covering the internal nappes are found at altitudes of 1000 to 1700 m, while in the Braşov basin the average elevations are around 500 m. These elevations vary from 530 m in the Târgu Secuiesc sub-basin, to 500 m in the Sfântu Gheorghe sub-basin and 460 m in the Bârsa-Baraolt sub-basin (Fig. 4.10b). The elevations of Lower Pleistocene sediments are important because they are considered to be deposited at same basin-level as sediments in Focşani foreland basin (i.e. similar Eastern Paratethys brackish/lacustrine fauna; Rögl, 1999; Marinescu and Popescu, 1995; Olteanu, 2003).

Inversion of the Pliocene extension and herewith exhumation in the Braşov basin have started close to the late Early Pleistocene when the Eastern Paratethys sedimentation was replaced by the fluvial deposition. The onset of the late Early Pleistocene uplift was tentatively chosen around 1.3 Ma, middle of Early Pleistocene (1.81-0.781 Ma) similar as in the Focşani basin (Subchapter 4.2.1). The rates of uplift were probably different in various places inside the basin as it is recorded by the present-day altitudes of the lacustrine sediments, which are found at ~1000 m south and at ~700 m north of the Bârsa-Baraolt sub-basin (Fig. 4.11), while near its centre they are found at 500 m.

If the present-day basin elevation of 500 m is subtracted from the elevations of 1000 and 700 m, it results in topographic differences of 500 and 300 m. These values represent the amounts of uplift recorded since the late Early Pleistocene in the southern and northern parts of the basin, respectively. In fact, the centre of the Braşov basin does not record uplift-related incisional features, geomorphological observations demonstrating that the fluvial sedimentation kept pace with the on-going subsidence (Posea, 1981). This second-order feature is juxtaposed over the generalised uplift affecting the SE Carpathians.

#### *4.6.2.1. Middle Pleistocene: upper strath-terraces*

Starting with the Middle Pleistocene, the river adjustment to the continuous deformation is reflected by the six terrace levels (TI-TVI; Fig. 4.2b), presently located between 100 and 10 m above the riverbeds (Fig. 4.3b). The oldest level TI placed at 100 m has an assumed Mindelian I age (Ghenea et al., 1971) equivalent to MIS13 (472-502 kyr; Table 4.3b). This allowed to estimate the amount of the previous late Early Pleistocene uplift, which took place from 1.3 Ma to either 0.7 Ma (end of Early Pleistocene) or 0.5 Ma (older age limit of level TI). The latter was chosen as the uplift limit because it corresponds to the formation time of this level. If the present-day elevation of terrace level TI of 100 m is subtracted from the present-day elevation of the Lowermost Pleistocene strata of 500 m in the south and 300 m in the north, it results in topographic differences of 400 and 200 m, respectively. These values correspond to the uplift taking place during the late Early Pleistocene. Tentative uplift rates have been calculated for a time period of 0.8 Ma and the values are 0.5 and 0.25 mm/yr, respectively (no error bars could be computed).

The Middle Pleistocene terraces belong to strath levels TI-TII (Fig. 4.2b), which are made up of sands, gravels and overlaying loess sequences with a total thickness of ~10 m (Table 4.1b). In the Vlădeni sub-basin, levels TI-TII are paired-type terraces developed equally on both banks of the Homorod river, gradually disappearing eastward towards the basin centre. In the Sfântu Gheorghe sub-basin, these levels are unpaired and poorly preserved on the right bank of the Olt river, which indicate that the subsequent river incision/uplift was localised northward and westward (Fig. 4.2b). Sediments of these two terrace levels have been assigned by Ghenea et al. (1971) to units I and II of Mindelian glacial (Table 4.3b), equivalent to MIS13 (472-502 kyr; cold) for TI and to MIS12 (440-

472 kyr; warm) for TII (Fig. 4.9). From terrace ages estimated at 487±15 and 456±16 kyr and bedrocks placed at 90 and 70 m, have been computed the river incision rates of 0.2±0.006 and 0.15±0.005 mm/yr, respectively (Table 4.3b).

#### *4.6.2.2. Late Pleistocene-Holocene: middle and lower mainly fill-terraces*

Levels TIII-TVI are widespread in the entire Braşov system and are fill-terraces (Fig. 4.2b). Level TIII is situated at ~60-65 m above the riverbed (Fig. 4.3b) and is made up of 10-15 m of alluvial gravels and overlying clayey loess (Table 4.1b). It is particularly well preserved near the Miclăuş locality. This level has formed during the Rissian I glacial (Ghenea et al., 1971) equivalent to MIS6 (130-195 kyr; Fig. 4.9). From its age estimated at 162±32 kyr and the bedrock placed at 50 m, an incision rate of 0.3±0.06 mm/yr can be calculated (Table 4.3b).

Level TIVa (26 m) is an unpaired strath-terrace that is locally preserved on the right bank of the Olt river (Fig. 4.2b), with a bedrock surface placed at 20 m above the actual riverbed. In the Ariuşd sampling site (TO3; Fig. 4.7a), the loess sequence observed at the top of level TIVa is composed of two paleosols of 0.3 and 0.7 m, respectively, (S0 and S1) and two loess horizons of 2 and 3 m in thickness, respectively (L1 and L2). The age of 123±11 kyr obtained at 5 m depth fits to MIsub-S5e (warm climate), while the basal age of 130±12 kyr corresponds to the lower denudational unconformity (lower limit of MIsub-S5e; Fig. 4.8). Consequently, this terrace level formed during MIS6/MIsub-S5e transition (cold/warm) and the paleosol S13 would correlate with MIS3, probably equivalent to paleosol S13, located in the foreland basin along the Putna valley. From this terrace formation age, a river incision rate of 0.15±0.01 mm/yr may be inferred (Table 4.3b).

Level TIVb (30-40 m) is a fill-terrace made up of basal gravels (8 m), a reddish paleosol S12 (0.7 m), a loess horizon L1 (~2.2 m) and actual soil S0 (0.3 m). The L1 horizon dated by IRSL in the Arini site has an age of 46±5 kyr at 2 m in depth (Fig. 4.7b). If this corresponds to MIS3, then the basal reddish paleosol is close to an age of 59±6 kyr or older (Fig. 4.8). It corresponds to MIS4/MIS3 transition (cold/warm climate). The IRSL age does not confirm the Rissian I age (130-195 kyr) reported by Ghenea et al., (1971) by means of geomorphology (Fig. 4.8). The associated river incision rate is in the order of 0.5±0.03 mm/yr (Table 4.3b).

The lowermost levels TV-VI are fill-terraces situated at 15-20 and 10 m above the riverbed (Fig. 4.3b). These are both paired terraces in the Sfântu Gheorghe and Vlădeni subbasins and unpaired terraces on the SE part of the Transylvanian basin (Fig. 4.2b). Terrace fills are made up of 10 m of gravels and sandy-clayey deposits and usually overly the Lower Pleistocene lacustrine sediments. Mammal fossils from levels TV-TVI of the Sfântu Gheorghe basin indicate Würmian I and III ages for these levels (Dumitrescu et al., 1970; Ghenea et al., 1971). Units Würmian I and III would therefore correspond to MIsub-S5a-d (73-111 kyr) and MIS2 (13-28 kyr; Fig. 4.8), which indicate warm-cold and cold climates. Terrace ages estimated at 92±19 and 20±7.5 kyr and their denudation surfaces placed roughly at 17 and 6 m enabled to calculate the river incision rates of 0.2±0.04 and 0.3±0.1 mm/yr, respectively (Table 4.3b).

## 4.7. Consequences from terrace ages and incision rates on vertical movements and periods of terrace formation

The Middle-Late Miocene exhumation of the Romanian Carpathians gradually divided the Paratethys basins, resulting in separation of the endemic Pannonian stage in the interior of the Carpathian chain. Subsequently, the brackish/lacustrine sediments of the Transylvanian basin were completely exhumed during the Pannonian (9 Ma). The post-11 Ma subsidence affecting the foreland basin, i.e. south of the Trotuş fault (Fig. 4.1b), lead to a gradual deposition and westwards extension of the post-collisional sediments over the SE Carpathian nappes (Leever et al., 2006), with variable depositional rates in the range of 0.6-1.5 mm/yr (Tărăpoancă et al., 2003; Vasiliev et al., 2004).

During the Middle Pontian, a massive desiccation took place from 5.96 to 5.33 (Messinian Salinity Crisis; Krijgsman et al., 1999), best visible in the front of the South Carpathians (Clauzon et al., 2005; Stoica et al., 2007; Leever, 2007), being driven by the massive sea-level drop in the Paratethys driver, the Black Sea (Gillet et al., 2007). This event, visible only in the distal condensed sequences of the Focşani basin (Stoica et al., 2007), might be responsible for the Pliocene enhanced exhumation recorded by the AFT data (5-6 Ma; Sanders et al., 1999; Merten et al., accepted). During the Pliocene, the foreland subsidence reached the internal SE Carpathian nappes (Leever et al., 2006), possibly enhanced by an extensional episode localised in the Braşov basin (Fielitz and Seghedi, 2005 and references therein). Because the post-11 Ma sediments are missing between the Braşov and the Focşani basins, one could argue on formation in the former of a suspended lake system (i.e. tectonic lake, senso Garcia-Castellanos, 2006). However, because the Braşov basin records common Eastern Paratethys brackish/lacustrine fauna, it must have been at the same basin-level as the foreland basin, the Uppermost Miocene-Pliocene sediments in between being removed by exhumation/erosion during the Quaternary.

The Late Pliocene exhumation recorded by the low-T thermochronology, southwards along the Buzău valley (2-3 Ma; Merten et al., accepted) was correlated with the uplift of the western flank of the Focşani basin, subsidence in its centre (Leever et al., 2006) and the shift in the Carpathians drainage network (Fielitz and Seghedi, 2005). This event marks the onset of the generalised inversion and led to the isolation and exhumation of the Braşov basin during the late Early Pleistocene, the basin being ultimately uplifted. This first-order feature exhumed the Carpathian hinterland and changed its sedimentation pattern from lacustrine to alluvial. It cannot be subdivided in time and is treated as a bulk event (Fig. 4.10). However, the second-order detailed features observed in the Braşov basin can be differentiated in time on the basis of river incision and sediment thickness.

### 4.7.1. Early Pleistocene

During the late Early Pleistocene, large scale uplift was recorded in the thin-skinned nappes between the Focşani and Braşov basins. The 9°ENE-ward monoclinal tilting of the Lowermost Pleistocene strata near the western flank of the Focşani basin resulted in a river incision rate of 1.5 mm/yr and by extrapolation a rate of 3.0 mm/yr is suggested in the Vrancea half-window (Fig. 4.10a). During the same period, the flanks of the Braşov basin have been also uplifted, recording rates of 0.5 mm/yr in the north and 0.25 mm/yr in the south, respectively. Subsidence reached a rate of 1.1 mm/yr in the centre of the Focşani basin, decreasing westward to 0.1 mm/yr in the centre of the Braşov basin.

It can noticed that the amplitude of vertical movements varied along the W-E profile, reaching its maximum in the central external nappes (Vrancea half-window) and decreased to the east in the foreland and to the west in the intramontane basin, respectively (Fig. 4.10a). This asymmetric pattern might be explained by local tectonic features, i.e. high-angle reverse faults placed underneath of the basement of the external nappes which underwent uplift (E), associated with normal faults which triggered subsidence westward in the internal nappes (Fig. 4.10c; Ştefănescu et al. 1988; Chalot-Prat and Gîrbacea, 2000; Fielitz and Seghedi, 2005) and alkaline magmatism (Seghedi et al., 2004).

### 4.7.2. Middle Pleistocene

Deposition of a regional loess sequence started in the late Middle Pleistocene (168±48 kyr, Rissian glacial; Fig. 4.9), covering the tilted Lowermost Pleistocene strata of the Focşani basin and the sedimentary cover of strath-terrace T1 of the Putna river (Fig. 4.2a). The oldest sublevel (a) of terrace T1 formed after the Putna river abandoned this surface due to vertical channel incision at a deeper level (sublevel T1b). Subsequently, sublevel T1a was covered by loess deposits. The 1.4±0.4 mm/yr inferred incision rate (Fig. 4.10) results in early Late Pleistocene ages for sublevels b and c (Table 4.3a). The abandonment of sublevels T1a and T1b took place during the Rissian glacial stage, a time when loess continued to accumulate eastward in the foreland basin as indicated by the absence of the paleosols on the right bank of the Milcov river. In this area, Rissian incision of the Milcov river into the Lowermost Pleistocene conglomerates attained a reduced rate of 0.3±0.08 mm/yr (Fig. 4.10a).

In the Braşov basin, the Olt river incision resulted in formation of strath-terrace levels TI-TII (Fig. 4.2b) during the early Middle Pleistocene, known as the Mindelian glacial (Fig. 4.9), corresponding actually to warm and cold climatic periods. The paired and unpaired terrace types potentially indicate that the western and northern areas experienced likewise uplift, while subsidence continued in the basin centre and in the south (Fig. 4.10a). The Mindelian incision rates of 0.2±0.006 and 0.15±0.005 mm/yr into the internal nappes are comparable with those recorded during the late Early Pleistocene (Fig. 4.10a). Abandonment of level TIII by the Olt river took place during the Rissian glacial, the river incised into the sedimentary filling of the intramontane basin at a rate of 0.3±0.06 mm/yr, coeval with formation of sublevel T1a in the most external nappe and with loess deposition in the foreland basin (Fig. 4.9). The geomorphology suggests that during this time, the Olt river had a different trajectory than the present-day one, this river entering in the Transylvanian basin through the Vlădeni coridor (Ghenea et al., 1971; Fig. 4.2b). The increased sedimentary flux from the internal orogenic nappe during the Rissian glacial has facilitated rapid alluvial filling of the basinal areas, which were still subsiding.

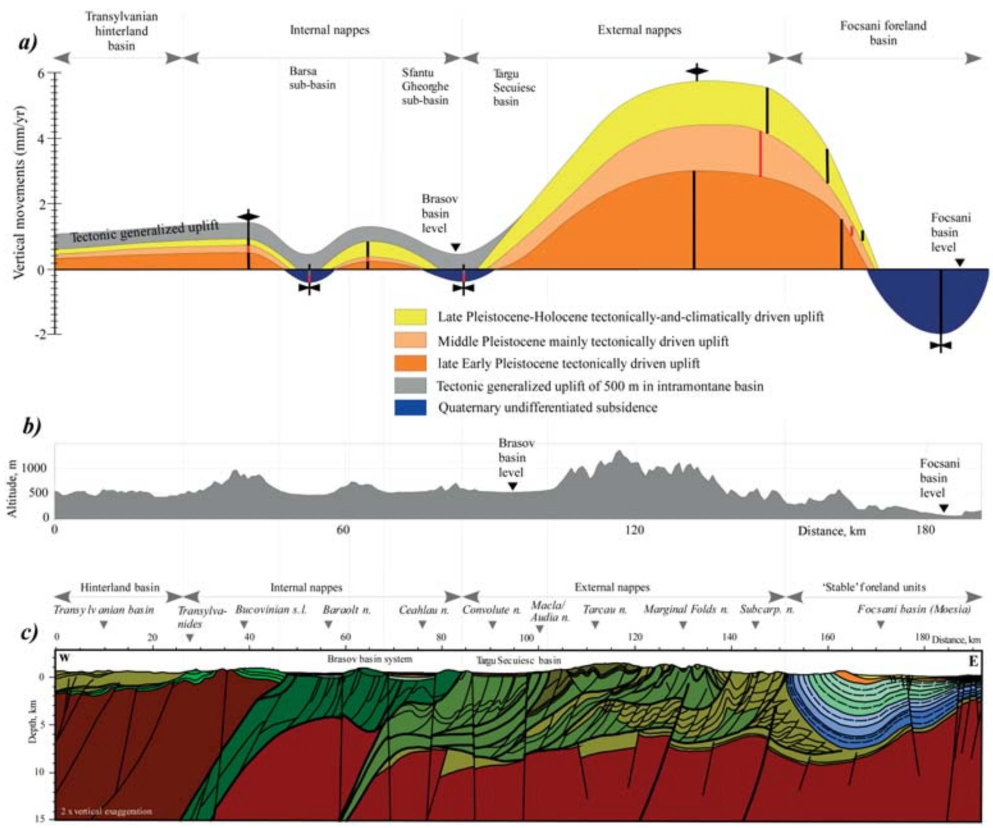
The various river incision rates might indicate that the earlier differential tectonic motions continued also during the Middle Pleistocene, equally affecting the orogen and the adjacent basins. However, the climatic contribution on river terrace formation should not be excluded as few levels were formed during full glacial stages (Rissian glacial). During this time, the Black Sea level dropped about 100 m (Winguth et al. 2000; Olteanu, 2004), having probably only a minor influence on river incision.

### 4.7.3. Late Pleistocene-Holocene

The river incision/uplift extended during the Late Pleistocene in the external nappes and the foreland basin, apparently increasing eastward and covering the previously subsiding areas, which were subsequently inverted and uplifted (Necea et al., 2005; Leever et al., 2006). The adjustments of the Putna and Milcov river channels to this deformation resulted in successive vertical incision events, corresponding to strath levels T2-T4, followed by short periods of aggradation reflected by the gravel deposits. Formation of these levels during either warm or cold climates of Würmian glacial (Fig. 4.9) indicates that the river incision was not restricted to any of these climatic conditions.

Deposition of the loess during the Würmian glacial covered large areas as far west as the Transylvanian basin. The loess sequence is well exposed in the foreland basin, i.e. on the left bank of the Putna valley, just opposite to the Țifeşti locality (Fig. 4.2a). Here, the

**Figure 4.10**




**Relative vertical movements recorded in the SE Carpathians during the Quaternary from Transylvanian to Focşani basins, with focus on two areas (black rectangles in Fig. 4.1b). River incision rates have been interpolated between the two areas, while the subsidence rates in the centre of the Focşani basin have been taken from Leever et al. (2006). (a) Quaternary asymmetric river incision/uplift rates across the orogen are represented by colours: dark blue stands for Quaternary undifferentiated subsidence, orange for the late Early Pleistocene tectonically-driven uplift, red for the late Middle Pleistocene-Holocene climatically-and-tectonically driven uplift, while light grey for tectonic generalized uplift of 500 m in the intermontane basin. Notice the highest amplitude of the uplift at the transition zone from the Carpathian orogen to the Focşani foreland basin. (b) Topography mirrors adequately the amplitude of deformation, lower in the internal nappes (W) and higher in the external nappes (E). Notice that the present-day level of the Braşov basin is higher with ~500 m relative to the foreland basin. Therefore, the river incision rates in the Braşov basin refer to the 500 m level. (c) Regional geological cross-section with the main structural units and adjacent sedimentary basins (see Fig. 4.1 for details).**

basal paleosol S13 unconformably overlying the Lowermost Pleistocene conglomerates, formed during a warmer period (MIsub-S5c?), subsequently followed by continuous deposition of loess through the Late Pleistocene (Würmian glacial). The Putna river incised into these sediments at a rate of 0.3±0.08 mm/yr (Fig. 4.10a). This rate is similar with the Middle Pleistocene one (Fig. 4.10a) indicating that the river incision remained constant in time. Note that the basal unconformity is a denudation unconformity which formed probably earlier, during the late Early Pleistocene.

In the Braşov basin, the amplitude of the river incision/uplift increased as well during the Late Pleistocene covering the basin flanks, while subsidence extended to the north towards the Baraolt sub-basin, accommodating a new drainage pathway of the Olt river (Fig. 4.2b; Ghenea et al., 1971). At the transition from Rissian glacial to Mindelian/Rissian interglacial, an increase of water discharge was recorded through widespread torrents incision. It resulted in formation of a local strath-terrace (level TIVa) due to the Olt river incision into the internal nappes, the associated incision rate was 0.15±0.01 mm/yr (Fig. 4.10a). Subsequently, this terrace was directly covered by continuous deposition of a loess sequence during the Late Pleistocene (Würmian glacial), except for a short warmer period when paleosol S13 formed (MIsub-S5c?), probably similar as in the foreland basin (Fig. 4.9). Abandonment of next level TIVb took place during the transition from warm to cold climate, the Olt river incised into the basin sediments at a slightly higher rate of 0.5±0.03 mm/yr (Fig. 4.10a). This pattern continued during the Latest Pleistocene. The lower levels TV-VI formed during the warm-cold and cold periods (Würmian glacial) coeval with the deposition of the upper parts of the loess sequences (Fig. 4.9), the associated river incision rates are 0.2±0.04 and 0.3±0.1 mm/yr, respectively (4.10a).

In the foreland basin, level T5 formed during transition from warm to cold climate and continued also during the cold period (Fig. 4.9), followed by the rapid loess deposition until the paleosol level S11, with variable accumulation rates. The 20-25 kyr age of the paleosol S11 overlying the gravels of the level T5, does not clearly fall into a warm climate. The Putna river incision rates associated to this level are gradually decreasing eastwards, from 1.5±0.2 to 1.3±0.13 mm/yr (Fig. 4.10a). Compared with the Middle Pleistocene, the rates are similar in the orogenic nappes, but increased significantly in the foreland basin, indicating a widening and eastward propagation of the uplifting areas. The younger levels T8-T11 formed during a warmer period (Holocene; Fig. 4.9) and indicate only few metres of incision, no rates were calculated.

Formation of levels T2 to T11 during either full climate (i.e. warm or cold) or climatic transitions (Fig. 4.9) indicates that river incision did not align preferentially to warm, cold or climatic transition. This suggests that both tectonics and climate controlled the river incision and terrace formation. During this time, the Black Sea level dropped about 60 to 70 m (Winguth et al. 2000; Olteanu, 2004), having probably no major contribution on river incision.

## 4.8. Conclusions

The river adjustment to the continuous Quaternary uplift recorded from the Focşani foreland basin in the east across the SE Carpathians to the Braşov intramontane basin system in the west, is documented by numerous terrace levels and tilting of the older strata. New terrace chronology, ages of terrace formation and loess deposition based on the new IRSL dating combined with the previous published data, have allowed to quantify the amplitude and the regional extent of the differential vertical movements. The foreland terrace levels document a higher cumulative uplift of ~250 m relative to the hinterland ones, where the uplift reached 100 m. During the late Middle-Late Pleistocene, the river incision and the associated uplift rates have reached the maximum amplitudes in the external orogenic nappes. The deformation amplitude decreased laterally towards the foreland and hinterland basins.

Formation of terrace levels T1 to T11 and TI to TVI during either full climate (i.e. warm or cold) or climatic transitions (Table 4.3) indicates that river incision did not fit preferentially to warm, cold or climatic transition. This suggests that both tectonic and climatic processes controlled the river incision and terrace formation in the SE Carpathians. During the Middle Pleistocene-Holocene, the Black Sea level dropped twice with 100 and 60 to 70 m, respectively, which might have played only a role in river incision and terrace formation.

The coupled subsidence in the foreland basin and uplift in the orogen, suggest an overall contraction. The accelerated uplifting episodes from the external nappes and the coeval subsidence from the foreland and intramontane basins might be explained by the deep-rooted high-angle reverse faults placed underneath in the basement underlying the Carpathian nappes. The previous research has inferred a late stage, out-of-sequence contractional phase affecting the SE Carpathians (e.g. Wallachian, Hipollyte et al., 1999 and references therein), which induced large-scale basement uplift beneath the Carpathians nappes (Landes et al., 2004; Bocin et al., 2005) associated with differential vertical movements and subsidence in the foreland (Leever et al., 2006), which can be the effect of fold-like geometries at shallow crustal scale (Cloetingh et al., 2004). To clarify/confirm one of these two mechanisms, more detailed research is required.