Tectonostratigraphic model for underfilled peripheral foreland basins: An Alpine perspective

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ABSTRACT

Advances in the development of quantitative models of foreland basin stratigraphy have outpaced the observational data used to constrain the input parameters in such models. Underfilled peripheral foreland basins comprise a broad threefold subdivision of depositional realms that translates into three stratigraphic units which are commonly superimposed during basin migration; these units are here termed the "underfilled trinity." The three units of the trinity reflect (1) carbonate deposition on the cratonic margin of the basin (the lower unit), (2) hemipelagic mud sedimentation offshore from the cratonic margin of the basin (the middle unit), and (3) deep water turbiditic siliciclastic sedimentation toward the orogenic margin of the basin (the upper unit). Theoretical predictions of how such a complex basin fill initiates and evolves through time are not currently available; hence this study reviews the stratigraphy of underfilled peripheral foreland basins and provides a unique data set comprising rates of thrust advance and basin fill migration for the Tertiary foreland basin of the European Alps.

The Paleocene to Oligocene Alpine foreland basin of France and Switzerland comprises a well-developed underfilled trinity that is preserved within the outer deformed margins of the Alpine orogen. Structural restorations of the basin indicate a decrease in the amount of basin shortening from eastern Switzerland (68%) to eastern France (48%), to southeastern France (35%). Structurally restored chronostratigraphic diagrams allow rates of basin migration to be calculated from around the Alpine arc. Paleogeographic restorations of the Nummulitic Limestone (lower unit) illustrate a radial pattern of coastal onlap on to the European craton. Time-averaged rates for northwestward coastal onlap of the underfilled Alpine basin across Switzerland were between 8.5 and 12.9 mm/yr. Time-equivalent westward to southwestward coastal onlap rates in France were between 4.9 and 8.0 mm/yr. The direction of migration of the cratonic coastline of the basin was parallel to the time-equivalent thrust motions, and oblique to the Africa-Europe plate motion vector. By comparing rates of thrust propagation into the orogenic margin of the basin to rates of coastal onlap of the cratonic margin of the basin, it is possible to suggest that the Alpine foreland basin of central Switzerland migrated with an approximately steady state geometry for at least 210 km northwestward over the European craton. The westward and southward decrease in the basin migration rate around the Alpine arc was associated with an increase in the degree of syndepositional normal faulting on the European plate; this is thought to relate to the opening of the Rhine-Bresse-Rhône graben system.

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INTRODUCTION

Peripheral foreland basins develop in response to the load of the thickened crust that results from continental collision (Dickinson, 1974; Beaumont, 1981; Allen et al., 1986). The sedimentary infill of foreland basins records the interaction between the growth of the thrust wedge, the isostatic adjustments of the cratonic lithosphere to thrust loading and additional bending moments, eustasy, and the surface processes that redistribute material from the mountain belt into the surrounding basins. Numerical models have tried to simulate the growth of thrust wedge–foreland basin systems, and attempted to evaluate the significance of individual parameters and how they interact (Jordan, 1981; Stockmal and Beaumont, 1987; Flemings and Jordan, 1989; Sinclair et al., 1991; Johnson and Beaumont, 1995). The challenge for field geologists studying foreland basins is to test the validity of these models, and to provide input on the rates and styles of the tectonic and surface processes that develop during the evolution of their basin. This study provides a synthesis of the tectonic and stratigraphic evolution of underfilled peripheral foreland basins from around the world, and then focuses on the most thoroughly documented of these: the Tertiary Alpine foreland basin of France and Switzerland (Fig. 1).

Numerous publications have demonstrated how peripheral foreland basins evolve from an underfilled to a filled or overfilled depositional state (Graham et al., 1975; Labaume et al., 1985; Covey, 1986; Homewood et al., 1986; Houseknecht, 1986; Tankard, 1986; Ricci-Lucchi, 1986; Grotzinger and McCormick, 1988; Coakley and Watts, 1991). The definition of the depositional state of a basin requires a reference frame within which the degree of filling can be defined. Computer-generated basins define the degree of filling of a flexural depression with reference to the point of zero lithospheric deflection on the cratonward margin of the basin (Flemings and Jordan, 1989). However, when studying the fill of ancient basins, locating this point is not always possible, with a few notable exceptions (DeCelles and Burden, 1992). Therefore, in ancient settings, the degree of filling of the basin is commonly approximated from the long-term trends in the sedimentary facies found in the basin; i.e., deep marine facies equate with underfilled, shallow marine–distal continental facies equate with filled, and fully continental facies equate with overfilled (Tankard, 1986; Homewood et al., 1986; Labaume et al., 1985; Sinclair and Allen, 1992). This use of sea level as the reference frame implies that mean sea level and the elevation of the stable craton do not differ significantly (within ~200 m) when considering the long term (>5 m.y.) development of the basin. The use of sedimentary facies as an indication of basin filling can only be applied to long-term trends in sedimentation, and facies that are used to identify an underfilled state must have been deposited in significant (>200 m) water depths.

The controlling factors on the degree of filling by siliciclastic sediments shed from mountain belts into their neighboring foreland basins have been assessed using quantitative models (Stockmal and Beaumont, 1987; Flem-

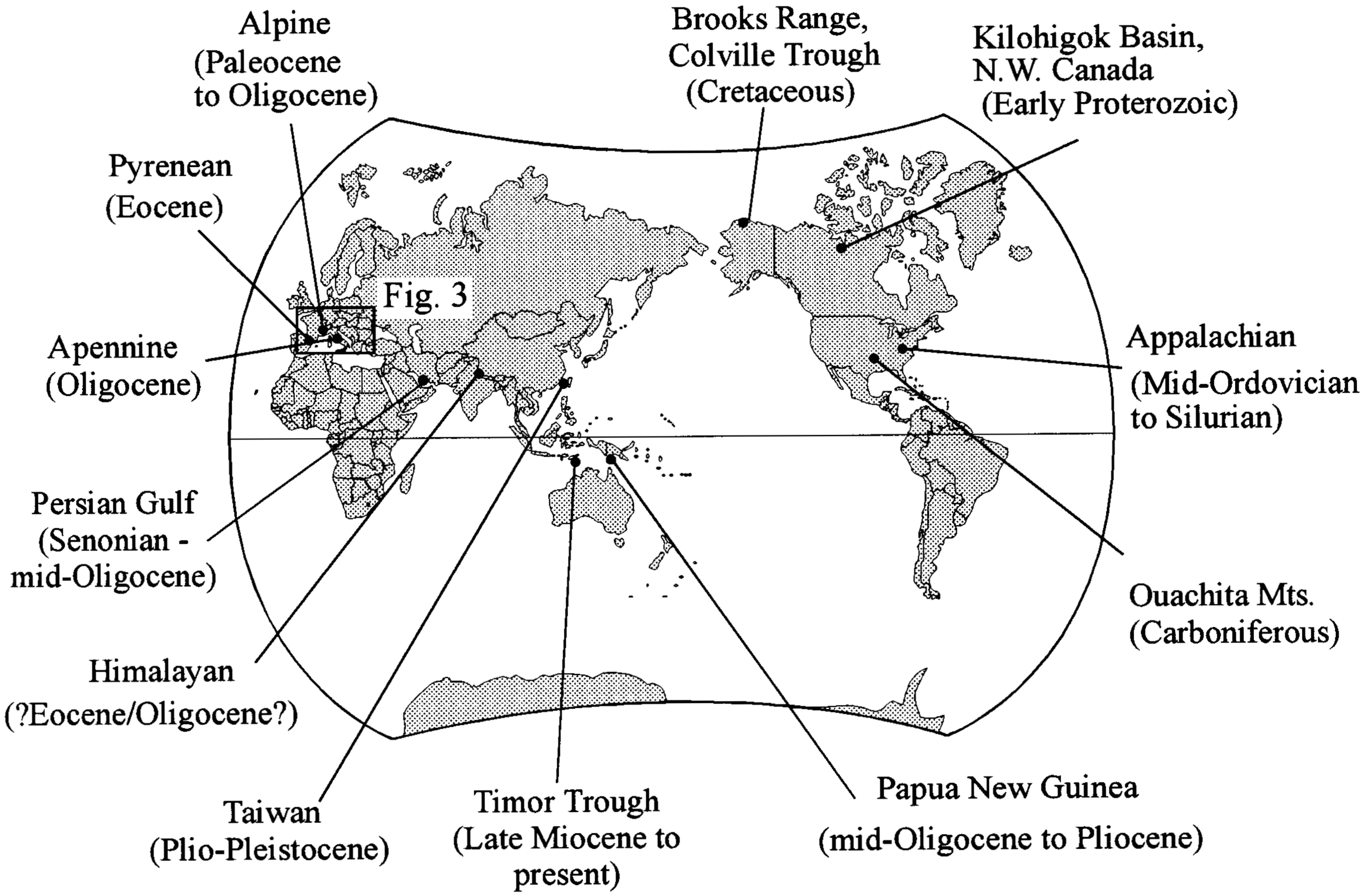

Figure 1. Map of the world showing locations of documented underfilled peripheral foreland basins. Box locates Figure 3. Data on the individual basins located on the map are given in Table 1.

ings and Jordan, 1989; Sinclair et al., 1991; Sinclair and Allen, 1992; Watts, 1992). The results of these models suggest three basic factors that encourage the transition from an underfilled to an overfilled depositional state: (1) slowing thrust wedge advance, (2) increasing exhumation and sediment production from the thrust wedge, and (3) increasing flexural rigidity of the underlying cratonic plate. Royden (1993) considered the characteristics of underfilled and overfilled peripheral foreland basins within the broader context of the forces applied to the flexed plate during collision.

The aims of this study are first to generate a synthesis of underfilled peripheral foreland basin sedimentation, and second to use the Alpine example to illustrate the links between rates of thrust front advance, and the rates and motions of basin migration as recorded by the sedimentary fill of the basin. This represents a unique data set that is able to combine detailed knowledge of the migration of basin facies on to the European craton with documentation of the timing of deformation in the associated thrust wedge.

UNDERFILLED TRINITY

Underfilled peripheral foreland basin stratigraphy has been documented from numerous basins worldwide, varying in age from Archean to the present day (Fig. 1). A review of the literature on underfilled foreland basins (Table 1) reveals a common stratigraphic signature that can be described in terms of three lithostratigraphic units (here termed "the underfilled trinity"; Fig. 2), which are markedly diachronous and commonly superimposed on top of one another during cratonward migration of the facies belts. This tripartite division of underfilled foreland basin stratigraphy has been recognized from individual basins such as north of the Alps (the "trilogie Priabonienne," Boussac, 1912) and by the variation in sedimentation rates in the Middle Ordovician of the Appalachian foreland basin (Shanmugan and Walker, 1980). The underfilled trinity can be summarized as follows (Fig. 2): (1) a lower unit that may or may not be underlain by a basal unconformity and that comprises a variable thickness (0–2500 m, average 550 m) of shallow marine limestones and sandstones; (2) a middle unit composed of 50–4000 m (average 830 m) of mudstones rich in pelagic microfauna; and (3) an upper unit comprising 45–4000 m (average 2000 m) of sandstones dominated by turbidites and classically termed "flysch."

This succession is commonly separated from the underlying passive margin strata by an unconformity (Veevers, 1971; Murris, 1980; Jacobi, 1981; Mussman and Read, 1986; Grotzinger and McCormick, 1988; Pigram et al., 1989; Allen et al., 1991; Coakley and Watts, 1991). some of these unconformities have been interpreted as the result of erosion over a topographic high generated by forebulge uplift of the foreland plate (Rowley and Kidd, 1981; Jacobi, 1981; Stockmal et al., 1986; Grotzinger and McCormick, 1988; Allen et al., 1991; Coakley and Watts, 1991; Waschbusch and Royden, 1992; Crampton and Allen, 1995).

Each unit of the trinity is described herein in terms of dominant facies types, depositional environments, and stratigraphic development; comparisons are made between basins of different ages and geographic setting

TABLE 1. SUMMARY OF THE MAIN CHARACTERISTICS OF THE UNDERFILLED PERIPHERAL FORELAND BASINS

Location	Age	Rate of thrust advance (mm/yr)	Te (km)	Unit thickness; facies; formation names			Paleoclimate	References
				Lower unit	Middle unit	Upper unit		
Alps, Switzerland	Paleocene to mid-Oligocene	6–12	<30	5–50 m; Nummulite-rich hemipelagic limestones underlain by localised conglomerates; Nummulitic Limestones	50–400 m; Gray calcareous mudstones; Globigerina Marls, Marnes Bleue	200–1000 m; Volcanic-rich, immature turbiditic sandstones; Taveyannaz, Champsaur, and Annot Sandstones, Val d'Illiez Formation	Subtropical	Matter et al. (1980); Homewood et al. (1986); Pfiffner (1986); Herb (1988); Caron et al. (1989); Allen et al. (1991); Sinclair and Allen (1992)
Appalachian and Taconic foredeep, eastern United States	Middle Ordovician	19–33	80–130*	0–300 m; Massive, light gray calcareous mudstone, pellet intraclast packstone, fenestrae; Mossheim/New Market Limestone (Virginia), Larrabbee Formation (New England), Lower Trenton Limestone (New York)	120–760 m; Graptolitic mudstones, black calcilutite turbidites, black shales dominate toward top; Utica Shale Formation (New England, New York), Paperville Formation (Virginia)	45–4000 m; Coarse-grainedsandstone turbidites, siltstones, and mudstones; Tourelle, Martinsburg, Cloridorme, Tellico and Knobs Formations	Tropical latitudes with cool ocean currents	Thomas (1977); Benedict and Walker (1978); Read (1980); Shanmugan and Walker (1980); Rowley and Kidd (1981); Cisne et al. (1982); Walker et al. (1983); Bradley andKusky (1986); Hiscott et al. (1986); Lash (1988); Bradley and Kidd (1991)
Apennines, Italy	Lower Oligocene to present	N.D.[†]	20±5	N.D.[†]	Hemipelagic, calcareous mudstones	3000 m; Sandstone turbidites and mudstones with interbedded hemi-pelagic sediments; Marnoso Arenacea and Macigno Formations	Warm temperate	Ricci-Lucchi (1986); Royden et al. (1987)
Brooks Range, North Slope, Alaska	Upper Jurassic to top Lower Cretaceous	N.D.[†]	65*	<80 m; Bioturbated quartzose, fine-grained sandstone/mudstones; Kemik Sandstone	60–90 m; Mudstones with localized beds of matrix supported pebbles; Pebble Shale	Olistostromes and debris flows to more distal turbidites and condensed mudstones; Okpikruak Formation, Hue Shale, Colville Group	Latitude 60°–70° north	Molenaar (1983); Hubbard et al. (1987); Coakley and Watts (1991)
Himalayan foredeep, north India, and Pakistan	Cretaceous to middle Eocene	N.D.[†]	80–100*	Nummulitic Limestones	N.D.[†]	Deep marine turbidites (little documentation)	Tropical to subtropical	Gansser (1964); Sahni and Kuhmar (1974); Graham et al. (1975)
Kilohigok basin, northwest Canada	Proterozoic (1.9 Ga)	0	8–12	2500 m; Vertically stacked shelf sandstones, mud-stones, and minor carbonates with internal unconformities; Hackett and Rifle Formations, Bear Creek Group	400 m; Mudstones; Hackett and Rifle Formations, Bear Creek Group	1000 m; Submarine fan and interfan turbidites; Hackett and Rifle Formations, Bear Creek Group	10°–30° latitude, humid/tropical	Grotzinger and McCormick (1988); Grotzinger and Royden (1990); Hoffman and Grotzinger (1993)

TABLE 1. (*Continued*)

Location	Age	Rate of thrust advance (mm/yr)	Te (km)	Unit thickness; facies; formation names: Lower unit	Middle unit	Upper unit	Paleoclimate	References
Ouachita Mountains, Arkoma basin, central United States	Pennsylvanian	N.D.†	N.D.†	<100 m; Fluvial and tidal channel sandstones, fine sandstone/mudstone bioturbated tidal flats, offshore carbonate banks; Spiro Sandstone	Dark gray mudstone, few massive, clean sandstone (10–50 m); Atoka Formation	Submarine fan turbidite system; Atoka Formation	N.D.†	Morris (1974); Mack et al. (1983); Houseknecht (1986)
Papua New Guinea	Upper Oligocene to Pliocene	N.D.†	N.D.†	1000–1200 m; Initially red algae/large foraminifera limestone assemblage followed by broad (500 km) rimmed carbonate platform; Darai Limestone	Pelagic carbonates and mudstones; Kera Formation	Few gravity flow deposits (not documented in any detail); base of Aure Group	Transition from temperate to tropical	Pigram et al. (1989)
Pyrenees, Spain	Paleocene to upper Eocene	>5.0 (using turbidite onlap rate)	8–30*	100–200 m; Shelf carbonates rich in large foraminifera; Guara Group, Alveolina Limestone Group, Ager Formation	200–500 m; Blue marls deposited in a slope environment; lower Figols Group, Sagnari Formation (eastern region)	1500–2000 m; channel and lobe sandstone/mudstone turbidites with axial flow; Hecho Group	Subtropical to tropical	Labaume et al. (1985); Mutti (1985); Mutti et al. (1988); Zoetemeijer et al. (1990); Burbank et al. (1992)
Taiwan	Pliocene (4 Ma) to present	70.0 (plate motion rate)	N.D.†	No shallow marine sedimentation; passive margin/foreland basin transition in deep water	Up to 4000 m; Bioturbated silty mudstone	Prodelta mudstones and thin sandstones, no significant turbidite sedimentation	Tropical/ temperate rainforest	Covey (1986)
Timor trough	Upper Miocene to present	75.0 (from 3–1.8 Ma), now at 5.0	N.D.†	30 m; Shelf calcarenite (3.0–1.8 Ma) with reefs developing on present Australian shelf margin (Sahul shelf)	>350 m; Nannoplankton-rich ooze, clay rich upward 150 m normal faults offset ooze	No significant input from orogenic margin (Timor trough)	Tropical rainforest	Veevers (1971); Veevers et al. (1978); Bowin et al. (1980); Audley-Charles (1986)
Zagros foredeep, Persian Gulf	Senonian to Oligocene	11–37	5–22*	Mixed clastic rocks with broad (500 km) carbonate platform containing evaporites	Foraminiferal (globotruncanids) mudstones	Sandstone and siltstone turbidites, conglomerates and olistostromes near thrust front	Hot/arid	Murris (1980); Snyder and Barazangi (1986)

Notes: Underfilled peripheral foreland basins that have been documented in the literature and located in Figure 1. Their stratigraphy has been subdivided into the three units of the trinity. The references are the main references used for the data. Starred values for the equivalent elastic thickness (Te) of the cratonic plate represent data gathered from modern gravity surveys; other values have been derived from the stratal geometries of the basin infill. Data are presented in alphabetical order.

†N.D. = not determined.

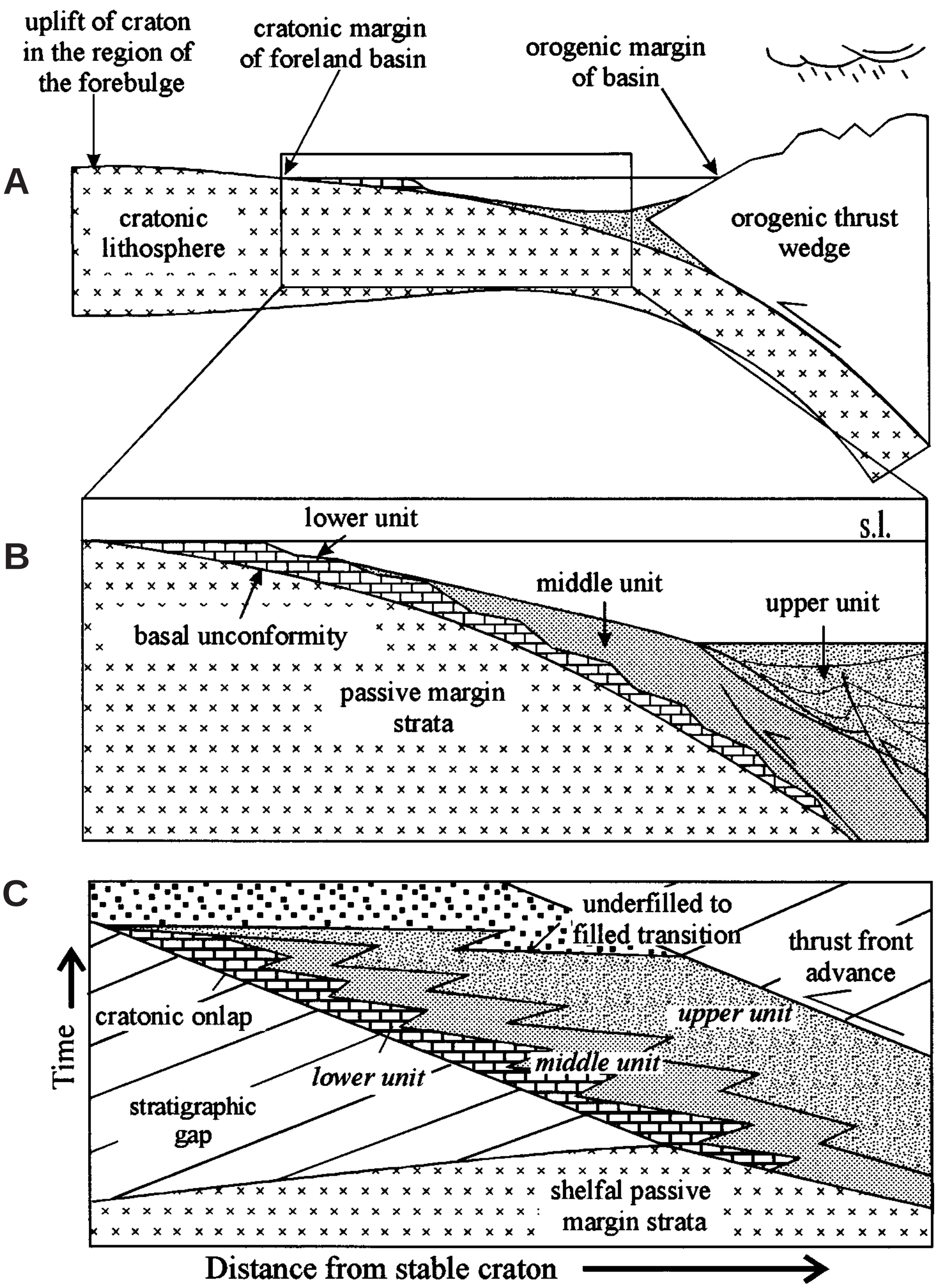

Figure 2. (A) Basinal setting of the underfilled trinity in a peripheral foreland basin setting. As the thrust load thrusts over the craton, so the foreland basin migrates at approximately the same rate. This results in a superposition of the three units of the trinity, as illustrated in B. The basal unconformity results from uplift and erosion of the craton while in the region of the flexural forebulge. (C) Chronostratigraphic representation of underfilled trinity shown in B. Cross-hatched areas represent lacunae. All three units are diachronous in response to the migration of the basin.

(Table 1). The controls on the development of these units and their tectonic significance are discussed after documentation of the Alpine example.

Lower Unit

The documented examples (Table 1) reveal that the lower unit of the trinity is dominated by carbonate platform deposits. Three exceptions to this trend are the Kilohogok, Brooks Range, and Ouachita basin sediments, where shallow marine siliciclastic deposits accumulated at the base of the succession. The carbonate platforms are typically of a ramp-type geometry (Dorobek, 1995), although the thickest (1200 m) carbonate deposits at the base of a foreland basin succession are from a rimmed platform from the Miocene Darai Limestones of Papua New Guinea (Pigram et al., 1989). In arid settings, the lower unit may contain evaporites, as recorded in the upper Campanian deposits of the Persian Gulf which compose a broad (500 km) platform with a central evaporitic pan (Murris, 1980).

The longer term development of the lower unit may show variations in the relative proportions of carbonate and siliciclastic sedimentation (see Alpine example below), and the type of platform may alter through time. For example, in the Oligocene to Pliocene deposits of Papua New Guinea a carbonate ramp dominated by red algae and large foraminifera developed into a coralline rimmed platform; this change in platform geometry is believed to have been associated with the transition from subtropical to tropical climates (Pigram et al., 1989).

Rapid along-strike variations in facies and thicknesses occur in the Alpine example (see below) and in the calcarenites of the Sahul Shelf south of Timor, where syndepositional normal faults with up to 150 m displacement cause abrupt thickness changes (Veevers et al., 1978).

Examples where shallow marine siliciclastic sedimentation dominates the lower unit reveal proximity to upland sources. The siliciclastic source of the Mississippian Spiro Formation of the Ouachitas was located in the Ozark dome to the north (Houseknecht, 1986). In the Kilohigok basin, the source was emergence and erosion of the Gordon Bay flexural arch, which was along strike from the main depocenter (Grotzinger and McCormick, 1988). In the Brooks Range, the source for the siliciclastics was the underlying Ellesmerian sequence (Molenaar, 1983).

Middle Unit

The middle unit of the underfilled trinity is dominated by mudstone rich in planktonic organisms. The mudstone may be very carbonate rich, e.g.,. foraminiferal ooze in the Timor trough (Veevers et al., 1978) or more silty and bioturbated, e.g., the Pliocene–Pleistocene of Taiwan (Covey, 1986). Generally, the middle unit is from 0 to 4000 m thick, and drapes the lower unit. Examples such as the Alps, Pyrenees, Papua New Guinea, Timor, and the Persian Gulf are dominated by carbonate pelagic-hemipelagic sediments that were deposited in deep water settings. The only example where significant sandstones are present in the middle unit is from the Arkoma basin; these are mature quartz sandstones, in contrast to the immature lithic-rich sandstones of the upper unit (Houseknecht, 1986). It is interpreted that feeder systems lateral to the basin generated submarine channels along the base of the downfaulted troughs along which these sands accumulated.

Biostratigraphic and paleobathymetric information is readily obtained from the middle unit. For example, in the Middle Ordovician of east Tennessee, various bathymetric indicators demonstrate that the top of the middle unit represents the time of maximum water depths in the basinal succession (Benedict and Walker, 1978); this is the same for the Alpine case described below.

Normal faulting is common during deposition of the middle unit, as described from the Arkoma basin. Normal faulting has also been documented from the middle unit of the Taconic foreland basin (Utica Shale) of New York (Bradley and Kidd, 1991), from the Pliocene oozes of the Timor trough (Veevers et al., 1978), and from the French Alps (see below). Thrust faulting during and immediately after deposition of the middle unit has been interpreted from the Alps (Elliott et al., 1985; Sinclair, 1992), Apennines (Ricci-Lucchi, 1986), and Pyrenees (Labaume et al., 1985; Mutti et al., 1988).

Upper Unit

The upper unit is dominated by thick successions of alternating turbiditic sandstone and mudstone that have been classically termed “flysch” (Beaudouin et al., 1970; Reading, 1986). The sandstones are characteristically highly immature and rich in lithic and volcanic detritus derived from erosion of the thrust wedge (Schwab, 1986). The maximum thickness for this unit comes from the Appalachian foreland basin, where as much as 4 km of submarine fan and basin plain sediments accumulated during Middle to Late Ordovician time (Hiscott et al., 1986). Transverse structural lineaments in the thrust wedge may act as conduits for sands and muds feeding the basin floor accumulations; this has been interpreted from the Alps (Lateltin, 1988) and Apennines (Ricci-Lucchi, 1986). Transverse structures within the South Pyrenean foreland basin separate shelf environments in the east from deeper water settings along the basin axis to the west (Mutti et al., 1988). Mutti (1985) suggested that within these deeper water deposits of the Hecho Group, the dominance of basinal lobe versus muddy channel-levee complexes varies in association with long-term relative sea-level fluctuations.

Thrust-faulted highs generated during deposition of the middle unit lead to ponding of turbidite flows of the upper unit, generating thick sandstone beds overlain by thick mudstone drapes (Pickering and Hiscott, 1985; Sinclair, 1992). Turbidites deposited in structurally confined basins commonly show evidence of reflection and deflection off containing slopes (Ricci-Lucchi and Valmori, 1980, Pickering and Hiscott, 1985; Sinclair, 1994).

Two modern examples of the upper unit from Taiwan and Timor contain little or no sand. In the Pliocene–Pleistocene of Taiwan, 4 km of offshore prodelta bioturbated silty mudstones are the only deep marine sediments; similar sediments are being deposited offshore Taiwan in the present-day foreland basin (Covey, 1986). The Timor trough contains only minor amounts of siliciclastic detritus delivered from the island of Timor (Veevers et al., 1978); sedimentation rates are ~0.4 mm/yr in the trough axis compared to ~0.1 mm/yr on the lower slope (Charlton, 1988).

Basinal Setting of the Stratigraphic Units

To summarize, the three stratigraphic units described above have been interpreted to reflect sedimentation in three regions of an underfilled peripheral foreland basin (Fig. 2). (1) The lower unit reflects shallow marine sedimentation on the cratonward margin of the basin. This unit accumulates on top of the underlying passive margin succession, and its basal deposits record the initiation of foreland basin subsidence at a given location. (2) The middle unit reflects sedimentation offshore from the cratonward margin of the basin. The initiation of the middle unit records the time at which relative sea-level rise linked to long-term flexural subsidence outpaced the growth of shallow water carbonate platforms (Dorobek, 1995). The upper part of the middle unit commonly represents the deepest part of the basin and hence records the location of the basin axis at a given time. (3) The upper unit is dominated by siliciclastic sedimentation derived from the thrust wedge, and accumulates at the toe of, and on top of, the thrust wedge in a manner similar to sedimentation in accretionary wedge settings.

ALPINE FORELAND BASIN

The Alpine foreland basin of France and Switzerland (Figs. 3 and 4) developed in response to continental collision between the African and Eurasian plates during early Tertiary time (Dewey et al., 1973). The underthrusting of the southward-facing Tethyan passive margin below the overriding African plate generated a submarine trench; this is considered here as the time of initiation of the Alpine foreland basin (Allen et al., 1991).

Paleogeographically, the narrow (50–100 km) marine trough of the early Alpine foreland basin was bounded to the south and east by the growing Alpine orogen, and to the north and west by the European craton. To the east, early Tertiary marine facies extend into Austria, and further into eastern Europe, following the outer edge of the Carpathians and possibly linking up with the Black Sea region through the region north of the Balkans (Ziegler, 1987).

During late Eocene and Oligocene time, western Europe was also affected by east-west extension, developing the north-northeast–orientated Rhine-Bresse-Rhône graben system (Fig. 3), which extends from central Germany south-southwestward through southeastern France to join the Mediterranean Sea at the Rhône delta (Bergerat et al., 1990).

The most visible expression of the Alpine foreland basin is the lowland area directly north of the Swiss Alps, which extends from Haute Savoie,

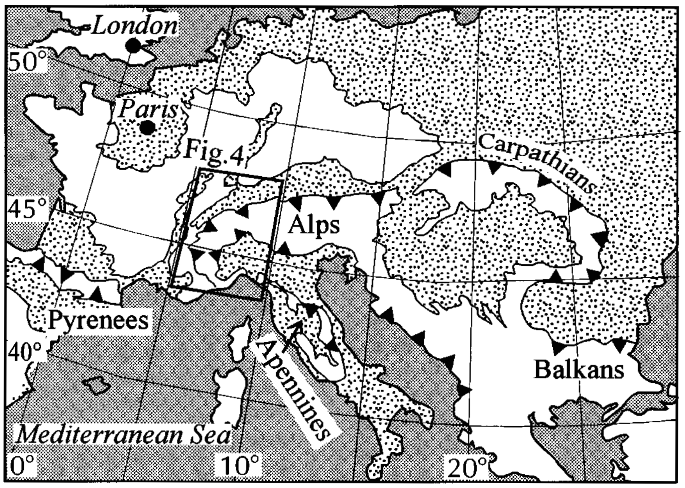

Figure 3. Distribution of Tertiary basins (stipple) and major mountain belts in Europe, with thrust fronts barbed. Figure 4 located in box.

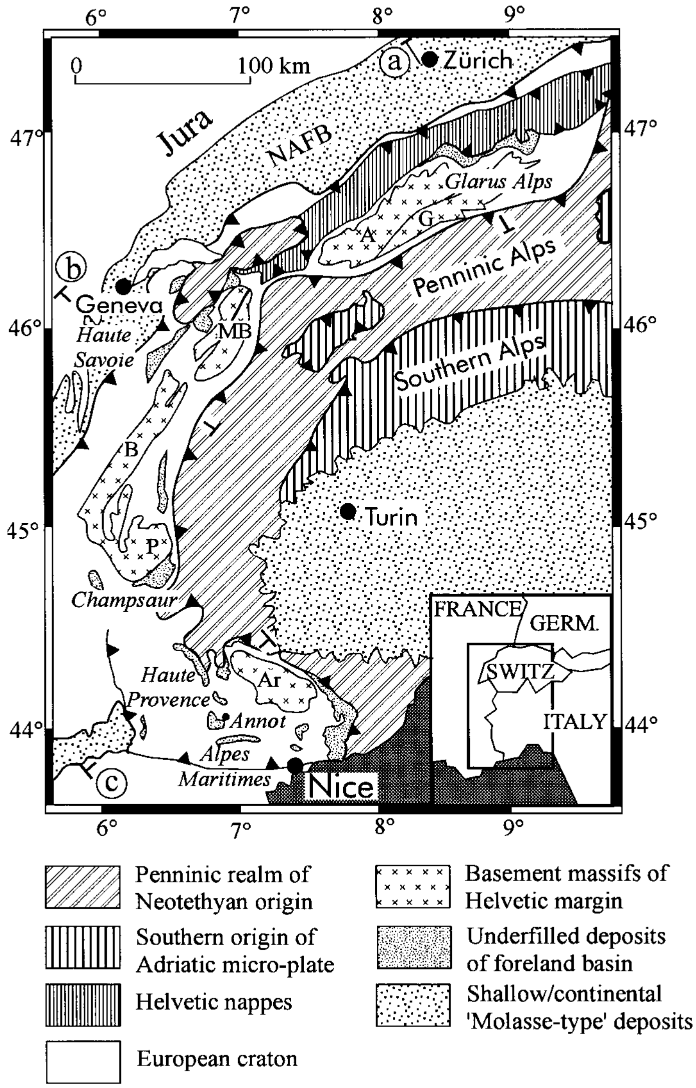

Figure 4. Simplified geological map of the western Alps, located in Figure 3. European basement massifs: G—Gotthard; A—Aar; MB—Mont Blanc; B—Belledonne; P—Pelvoux; Ar—Argentera; NAFB—North Alpine foreland basin. Cross sections shown in Figure 6 are located at a, b, and c.

France, in the west, through Switzerland to the Linz-Vienna area of Austria in the east (Fig. 4). South from Haute Savoie, the foreland basin of the French Alps is less obvious owing to subsequent Oligocene–Miocene uplift and deformation associated with the Bresse and Rhône graben systems.

The Mesozoic passive-margin succession is separated from the overlying foreland basin succession by a marked unconformity. In Switzerland, this unconformity has been interpreted to have developed in response to uplift and erosion of a flexural forebulge in advance of the Alpine thrust load (Allen et al., 1991; Crampton and Allen, 1995). Overlying this unconformity, the stratigraphic infill of the Alpine foreland basin has been well documented in Switzerland (Matter et al., 1980; Homewood et al., 1986; Berger, 1992) and in France (Beaudouin et al., 1975; Elliott et al., 1985; Allen and Bass, 1993). Sedimentologically, the basin fill can be divided into two stages; a Paleocene to mid-Oligocene deep marine (flysch) stage followed by a mid-Oligocene to late Miocene shallow marine and continental (molasse) stage. The focus of this study is the early deep marine stage, which reflects the underfilled stage of the basin's development (Allen et al., 1991; Sinclair and Allen, 1992).

Structural Setting of the Alpine Underfilled Basin

The predominantly deep water sediments of the underfilled succession can be found in the deformed outer margin of the Alpine thrust belt of France and Switzerland (Fig. 4). In Switzerland, the Paleocene to lower Oligocene remnants of the underfilled basin are now located in the Helvetic Alps (Figs. 4, 5, and 6A) and are characterized by stacked thrust sheets overlying a deformed, but not detached, cover to the external basement massifs (Trümpy, 1980; Pfiffner, 1986). It is within this deformed cover that the most extensive exposures of the underfilled trinity in Switzerland are found (Fig. 5; Matter et al., 1980; Pfiffner, 1986; Herb, 1988; Allen et al., 1991).

Farther west on the border of France and Switzerland (Fig. 6B), the underfilled trinity is preserved within the detached Helvetic nappes of Switzerland and in footwall synclines in Haute Savoie (Lateltin, 1988; Butler, 1989; Guellec et al., 1990; Mugnier et al., 1990). Farther south, extensive exposures of early flysch sediments are exposed in thrust sheets around the eastern margins of the Pelvoux Massif (Barbier, 1956; Deharveng et al., 1987) and in the relatively undeformed cover of the Pelvoux Massif at Champsaur (Waibel, 1990).

The least-deformed and best-preserved remnants of the underfilled basin can be found in the regions of Haute Provence and Alpes Maritimes in southeastern France (Fig. 4). The cross section through this area (Fig. 6C) illustrates the thin-skinned fold and thrust geometries generated by the Alpine deformation (Lemoine, 1972; Elliott et al., 1985). The Tertiary infill of the foreland basin is preserved in synclines perched on top of these structures. In this area, the Pyrenean orogeny extended its influence from Late Cretaceous to early Eocene time, generating east-west–orientated folds that are unconformably overlain by the Tertiary succession (Goguel, 1936; Lemoine, 1972; Siddans, 1979).

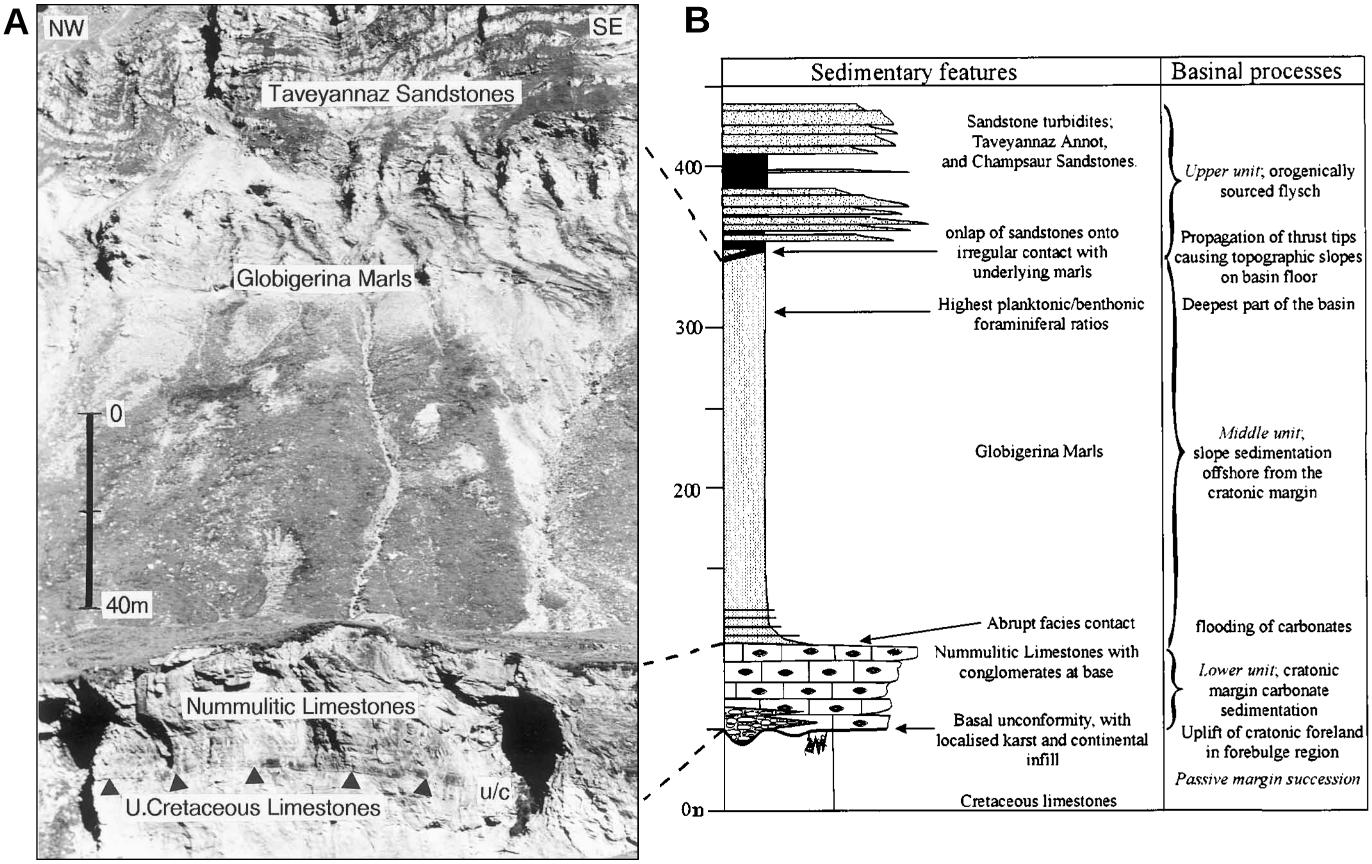

Figure 5. (A) Underfilled trinity of the upper Eocene–lower Oligocene of the Alpine foreland basin as seen at Panixerpass, Glarus Alps, eastern Switzerland (Sinclair, 1992). This section is from within the Infrahelvetic zone and shows folding and cleavage development, and is overthrust by the Helvetic nappes (Milnes and Pfiffner, 1977). The massive sandstone beds of the Taveyannaz Sandstones have been interpreted as the result of ponded turbidity currents in structurally confined piggyback basins (Sinclair, 1992). (B) Schematic graphical representation showing the main sedimentary features of the underfilled trinity from around the Alpine foreland basin. Dashed tie lines indicate where the boundaries between the three units of the trinity are located on the photograph from Panixerpass, Glarus Alps (B).

Palinspastic Restorations of the Underfilled Basin Stratigraphy

The three structural cross sections from around the external Alps (Fig. 6) are modified from previous sections that have been balanced for restoration purposes (Lemoine, 1972; Naef et al., 1985; Graham, *in* Elliott et al., 1985; Pfiffner, 1986; Pfiffner et al., 1990; Guellec et al., 1990; Mugnier et al., 1990). The construction of these sections is based on a combination of published maps, structural mapping, and seismic data. The sections have been restored using a line-length approach, and hence give minimum values for shortening. Nonplane strain deformation with out-of-the-section volume changes would increase the amount of total shortening. Values for these possible error bars are difficult to quantify, however, Hossack (1978) suggested that in the Caledonides of Norway, nonplane strain deformation may have accounted for 15% additional shortening through the section. Given the similarity in the mechanisms and amounts of shortening in the external Alps to the Norwegian Caledonides, a sliding scale of error was used with 15% error bars on the restored sections for the portions most distant from the pinline, reducing linearly to zero at the pinline.

The amount of shortening of the underfilled foreland basin of the Alps varies markedly along strike. The maximum value comes from eastern Switzerland (Fig. 6C), where a present-day distance of 53 km can be restored to 165 km, indicating 68% shortening. In Haute Savoie (Fig. 6B), 43 km of section can be restored to 82 km, indicating 48% shortening, and in southern France (Fig. 6A) 99 km can be restored to 153 km, indicating 35% shortening. Overall, these values indicate a decrease in shortening of the underfilled basin from eastern Switzerland to eastern France and an even lower value in southern France.

Inception of the Underfilled Basin

The inception of foreland basin subsidence and sedimentation is difficult to date precisely owing to incomplete preservation of the earliest deposits in the deformed external Alps. The inception of the foreland basin is defined as the earliest time at which there is evidence of increased subsidence prior to the arrival of orogenically derived sediment on the outer fringes of the European continental margin. The earliest well-documented evidence of this style

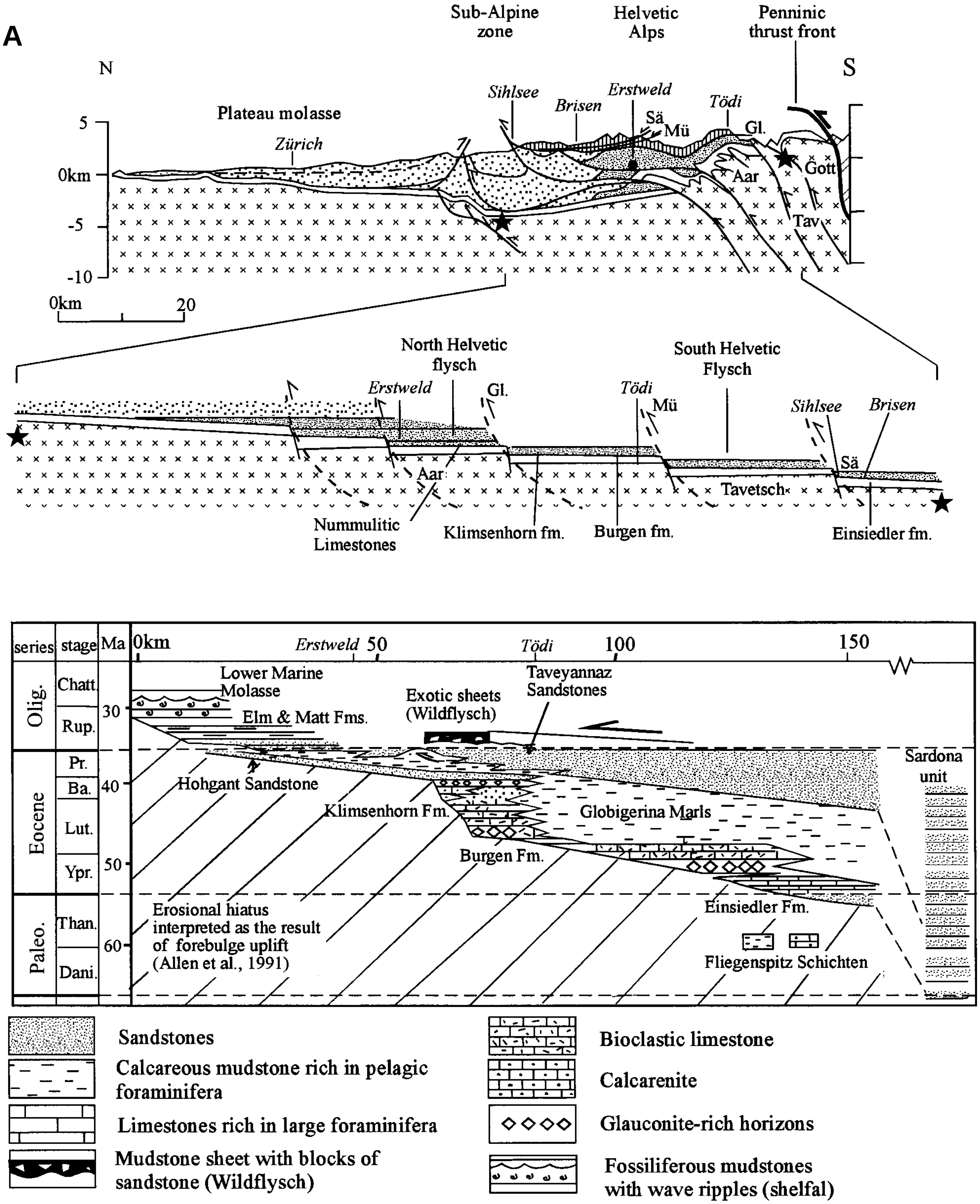

Figure 6. Balanced and restored cross sections of the underfilled portion of the Alpine foreland basin from three localities (A, B, C) around the Alpine arc (Fig. 4). Structural data from Pfiffner (1986); Pfiffner et al. (1990); Naef et al. (1985); Guellec et al. (1990); Mugnier et al. (1990); Butler (1989); Lemoine (1972); Elliot et al. (1985). The stars represent the pinline in the basement. The sections have been restored using a line length approach within the Jurassic stratigraphy. The restored sections act as a template for the chronostratigraphic diagrams at the bottom of each section. The stratigraphic information was compiled from Herb (1988); Pairis and Pairis (1975); Wegmann (1961); Charollais et al. (1980); Lateltin and Muller (1987); Campredon (1977).

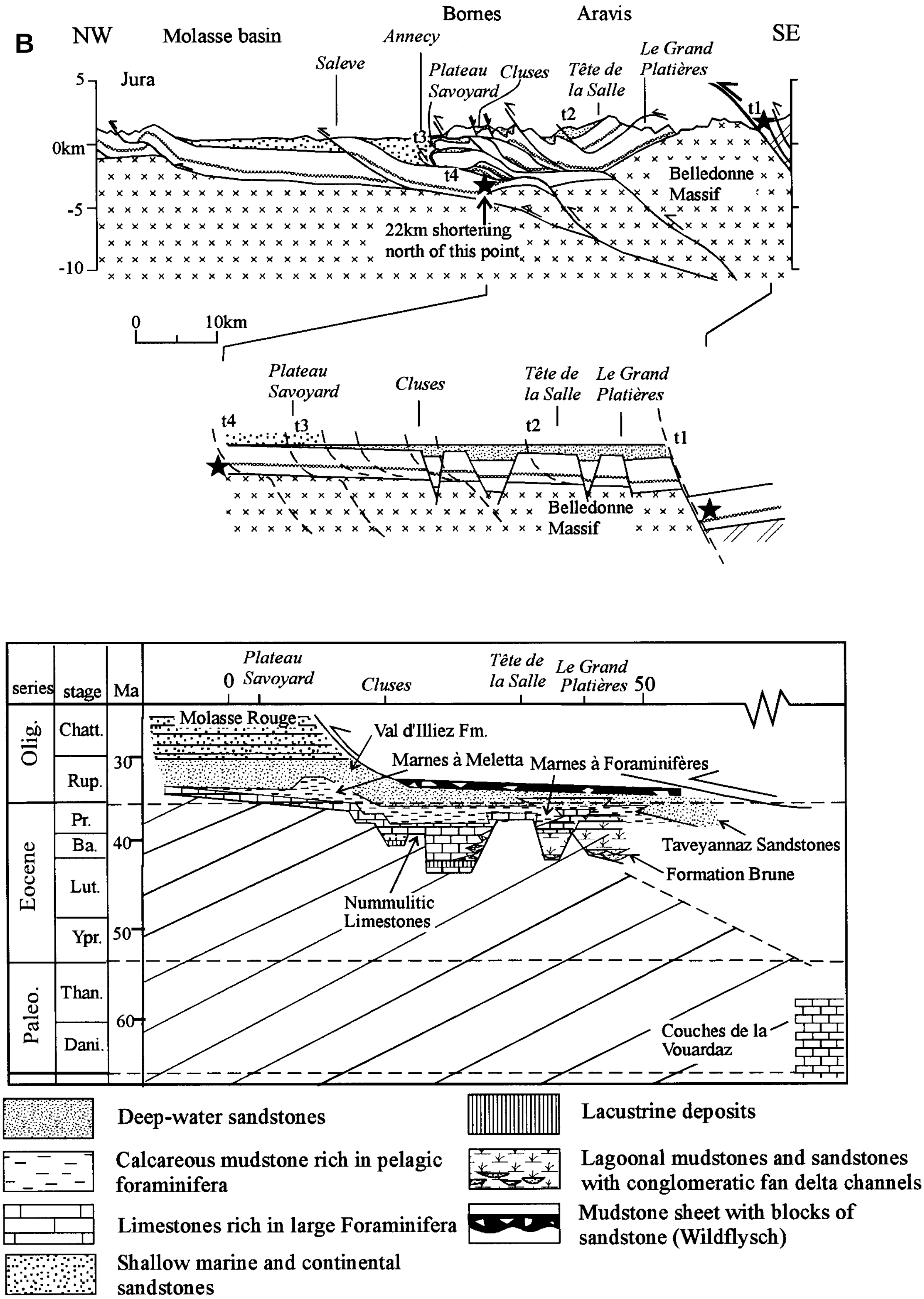

Figure 6. (*Continued*).

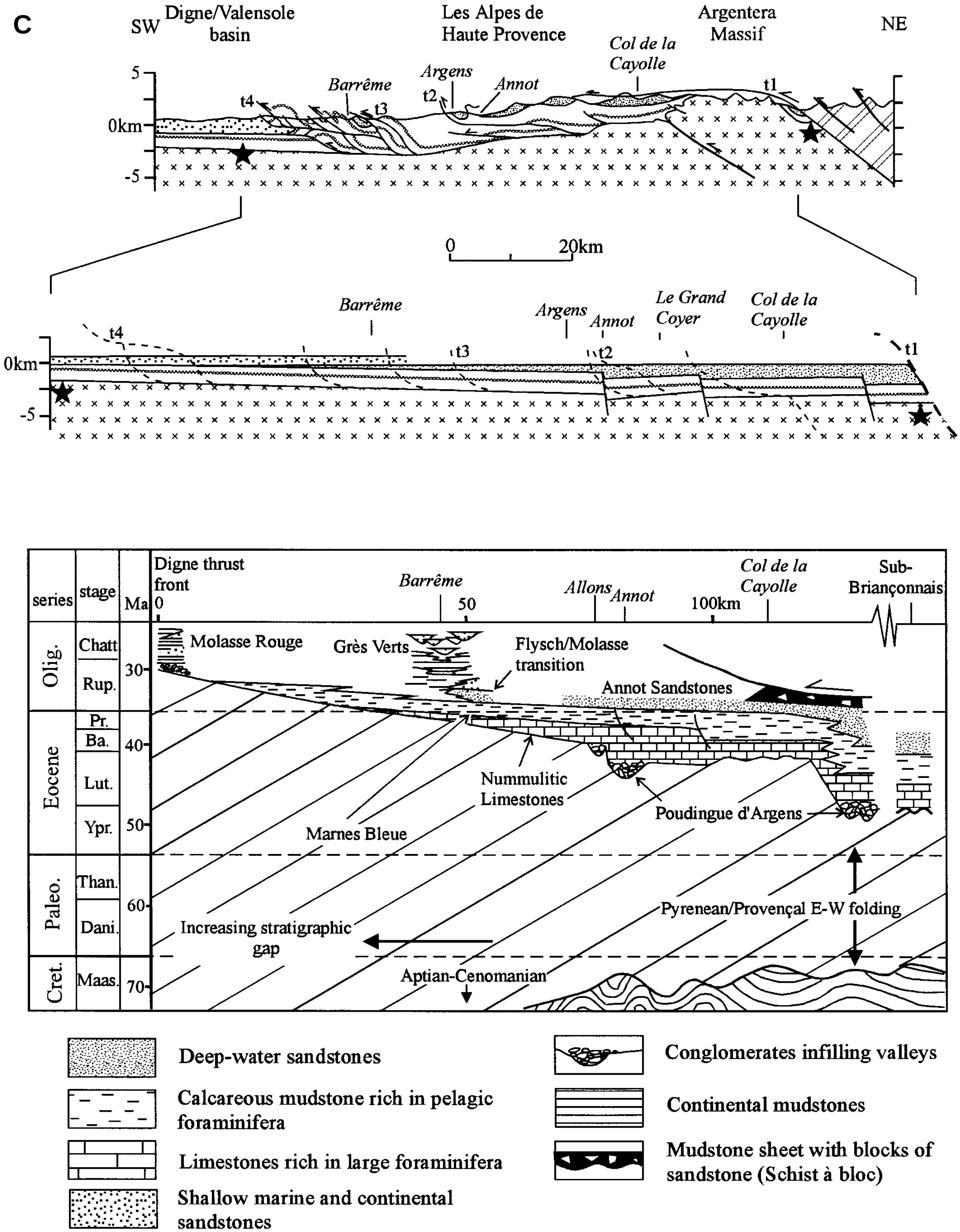

Figure 6. (*Continued*).

of sedimentation comes from the upper Paleocene– lower Eocene Blattengrat flysch unit of eastern Switzerland which was deposited on the outer passive margin (Herb, 1988; Lihou, 1995). The Maastrichtian and Paleocene Sardona flysch (Wegmann, 1961) of the very distal parts of the European margin contains flysch deposits that have been suggested to represent the initiation of foreland basin subsidence (Lihou and Allen, 1997). However, given that the source of these deposits was not the thrust wedge to the south, but localized fault-related highs exposed during sea-level lowstands, the interpretation for basin subsidence is speculative. Therefore, for the purposes of this paper, the time period of unequivocal underfilled foreland basin development is treated as late Paleocene to mid-Oligocene (Sinclair and Allen, 1992).

The initiation of load-induced flexure at a given location may occur earlier than the timing of increased subsidence prior to the accumulation of orogenically derived siliciclastics. Theoretically, there should be a small amount of subsidence in a location cratonward of the forebulge in the backbulge depozone (DeCelles and Giles, 1996). As yet, evidence for a backbulge depozone is not available from the Alps.

Stratigraphy of the Underfilled Basin

Detailed below is the upper Paleocene to mid-Oligocene underfilled trinity from the Swiss and French Alps. A graphical representation of the Alpine trinity is shown in Figure 5. The descriptions are based upon three chronostratigraphic charts from around the Alpine arc (Fig. 6) that show time against the paleogeographically restored locations for the measured sections.

Lower Unit. *(1) Central Switzerland (Fig. 6A)*. In central Switzerland, the lower unit comprises a deepening-upward succession of shallow marine foraminiferal limestones and glauconitic and/or phosphatic sandstones, which together are termed the Nummulitic Limestones; they rest unconformably on gently dipping Cretaceous strata (Fig. 5). The age range for the Nummulitic Limestones is early Paleocene (Fliegenspitz Schichten) to late Eocene (Hohgant Sandstone). Beneath the Nummulitic Limestones of the north and western part of the region are localized deposits of "siderolitique," which represent reddened karstic infill by terrestrial deposits; in some cases these penetrate 100 m below the unconformity surface (Herb, 1988).

The Nummulitic Limestones are typically 20–50 m thick, and show a long-term younging toward the north and west (Fig. 6A) associated with transgression over the European craton (Herb, 1988; Allen et al., 1991). In this study, the Nummulitic Limestones include a number of more localized members that reflect variations in the depositional settings on the cratonic margin of the basin.

Herb (1988) identified seven transgressive-regressive cycles that developed during deposition of the Nummulitic Limestones in Switzerland. The facies within these cycles have been interpreted to represent a storm-influenced carbonate ramp with nummulite-rich banks (Crampton, 1992). The Nummulitic Limestones of central Switzerland show only minor evidence of syndepositional faults controlling facies.

(2) Swiss/French border (Fig. 6B). In western Switzerland and Haute Savoie the Nummulitic Limestones are predominantly late Eocene and are much more variable than in central Switzerland owing to extensive syndepositional normal faulting (Pairis and Pairis, 1975; Lateltin and Müller, 1987; Villars et al., 1988). Facies and structural analysis reveal that fault-block highs were sites where coralgal reefs accumulated; these reefs shed material into the intervening troughs where finer sands and muds were deposited. Some of these faults are inferred to have extended perpendicular to the thrust front (i.e., northwest), preserving the older remnants of the lower unit in their hanging walls (Villars et al., 1988). Other faults were parallel to the subsequent thrust front (northeast) and accumulated channelized conglomerates and lacustrine deposits of late middle Eocene age in topographic lows that were subsequently blanketed by transgressive nummulite-rich limestones (Pairis and Pairis, 1975).

Stratigraphic thicknesses for the Nummulitic Limestones of this region are variable (Pairis and Pairis, 1975), but can be as much as 130 m, as documented from the Platé area, Haute Savoie (Fig. 4).

(3) Southern French Alps (Fig. 6C). This region of the southern French Alps includes the area to the south of the Pelvoux Massif (Fig. 4), where 50 m of upper Eocene Nummulitic Limestones were deposited on top of upthrusted basement and Triassic rocks (Fabré and Pairis, 1984; Fabré et al., 1985; Crampton, 1992). This is the only area where there is preserved evidence of significant amounts of Alpine shortening prior to the deposition of the Nummulitic Limestones and hence prior to the initiation of the foreland basin; this represents a significant anomaly in terms of the basin's development (see below). Below the Nummulitic Limestones of the Champsaur region are incised valleys which cut into Hercynian basement, Triassic and Jurassic strata, and contain breccias, conglomeratic red beds, and paleosols (Gupta, 1994). The valley fills are 100–500 m wide and 50–100 m deep. The clasts within the breccias and conglomerates are predominantly of local basement compositions.

Farther south in the regions of Alpes Maritimes and Haute Provence, valley-fill deposits are also present below the Nummulitic Limestones; these overlie broad erosion surfaces cut into the underlying Upper Cretaceous strata, infilled with as much as 200 m of channelized conglomerates (Poudingue d'Argens) and overbank sandstones and mudstones (Elliott et al., 1985; Thome, 1987). The conglomeratic clasts within these deposits are almost entirely derived from the underlying Upper Cretaceous strata. These fluvial to coastal conglomerates are blanketed by 5–50 m of transgressive Nummulitic Limestones with typical ramp-type facies as documented in Switzerland. Thickness variations within the Nummulitic Limestones indicate syndepositional normal faulting (Pairis, 1988). Elliott et al. (1985) linked the faulting to the early propagation of thrust tips into the basin, and the generation of lateral transfer structures.

Middle Unit. The middle unit around the Alpine arc is dominated by light gray calcareous, foraminiferal mudstones (Fig. 5). The formation name generally given to this unit is the Globigerina or Blue Marls. The Globigerina Marls are diachronous (Fig. 6A); their broadest age range is documented from central Switzerland, where southerly localities record dates of early to middle Eocene, and the more northerly localities range up to top Eocene (Herb, 1988).

The contact with the underlying Nummulitic Limestones ranges from an abrupt to very gradual transition from Nummulitic wackestones to calcareous mudstones. The Globigerina Marls range in thickness depending on their position relative to syndepositional faulting. The maximum thickness of 400 m comes from extensional hanging-wall strata, near Annot, Alpes Maritimes (Fig. 4; Pairis, 1988). As with the lower unit, the influence of syndepositional faulting is less well developed in Switzerland, and increases toward the French border. In southern France, the 350 m offset of the St. Benoit fault near Annot occurred during or just prior to deposition of the Globigerina Marls (Elliott et al., 1985; Pairis, 1988).

Micropaleontological studies of the Globigerina Marls from around the Alps (Eckert, 1963; Mougin, 1978; Charollais et al., 1980; Pairis, 1988; Herb, 1988) have identified high planktonic to benthonic foraminiferal ratios in the marls, and have been used to interpret depositional water depths at 500–1000 m (epibathyal and bathyal). Detailed analyses of foraminiferal ratios from the Globigerina Marls of the Annot area, Alpes Maritimes, have indicated that water depths increased upward through the marls and reached a maximum near the base of the overlying sandstones of the upper unit (Mougin, 1978). Hence, the transition from the middle to the upper unit is likely to record the maximum water depths in the basin.

Upper Unit. The upper unit composes much of the classic flysch deposits of the Alps (Homewood and Caron, 1982; Homewood and Lateltin, 1988). The names given to this formation vary around the arc, but the most common include the Taveyannaz Sandstones, the Val d'Illiez Sandstones, the Champsaur Sandstones, and the Annot Sandstones. The source for the sandstones of the Pelvoux region and north and east into Switzerland was the thrust wedge. The sandstone contains a high proportion of andesitic volcanic detritus (De Quervain, 1928; Vuagnat, 1952, 1983; Waibel, 1990). From latest Eocene into early Oligocene time, the composition of these sandstones became less andesitic and more ophiolitic. Farther south, in Alpes Maritimes and Haute Provence, the Annot Sandstones have a more variable composition derived primarily from surrounding basement and its cover (Ivaldi, 1974), although andesitic volcanic material is also present at certain horizons (Vernet, 1964; Stanley, 1980).

The transition from the calcareous mudstones of the middle unit to the siliciclastic mudstones and sandstones of the upper unit can be abrupt or gradational. In the Annot Sandstones of Haute Provence, the contact is a downlap surface reflecting progradation of the turbidite system over the Globigerina Marls in the deepest part of the basin.

Stratigraphic thicknesses for the upper unit sandstones vary from 50 to 2000 m depending on their depositional setting with respect to syndepositional emergent thrust tips (Sinclair, 1992), and on the rate at which the depocenters were overthrust by the encroaching thrust wedge. The age of the sandstones is predominantly late Eocene and basal Oligocene (Fig. 6). If we include the flysch deposits of more southerly origin which represent the upper unit of the very earliest stages of basin development (Sardona flysch, Blattengrat flysch, sub-Brianconnais, etc.), then we can consider the oldest deposits to be of late Paleocene to early Eocene age (Fig. 6).

The Taveyannaz Sandstones of Switzerland are dominated by basin plain and lobe turbidite deposits (Siegenthaler, 1974; Lateltin, 1988; Sinclair, 1992). In the Glarus Alps of eastern Switzerland (Figs. 4 and 5) the Taveyannaz Sandstone turbidites accumulated in two structural depocenters, a perched piggyback basin where topographic confinement caused ponding of the flows (Fig. 5), and a trench-floor setting, the trench parallel to eastward paleoflows (Sinclair, 1992). A similar interpretation of piggyback basinal settings for Taveyannaz Sandstone sedimentation is given for western Switzerland and Haute Savoie (Lateltin, 1988). Farther south in the Alpes Maritimes and Haute Provence, full depositional systems have been reconstructed for the Annot Sandstones; i.e., deltaic feeder systems transported material northward from the Corsico-Sardinian Massif into a deeper water turbidite system (Stanley, 1975; Elliott et al., 1985; Jean, 1985; Ravenne et al., 1987). In this area, sediment was also delivered from the evolving thrust belt to the east (Ivaldi, 1974), and recent reinterpretations of some of the more easterly outcrops have suggested a shallow marine source area feeding material to the northwest (Sinclair, 1993), which is in agreement with previous provenance studies (Stanley, 1961; Ivaldi, 1974).

Upper Eocene Paleogeography and Coastal Onlap

Using the restored chronostratigraphy from around the Alpine arc (Fig. 6), it is possible to construct a paleogeographic map of the basin for a particular time slice such as the late Eocene–early Oligocene (Fig. 7). Such a reconstruction uses the three restored lines obtained from this study as anchor points, and links them together using previous reconstructions (Debelmas, 1975; Kerkhove, 1980; Herb, 1988). This shows the spatial relationship between the three depositional realms within the basin, and the location of contemporaneous faulting.

A reconstruction of the position of the Nummulitic Limestone carbonate ramp through time records the position of the cratonic margin of the foreland basin (Fig. 8). This permits an evaluation of the pattern and rates of migration of the cratonic margin of the basin for this time interval. The pattern is broadly radial with northwestward migration in Switzerland to west to southwestward migration in the south of France. This radial migration corresponds almost exactly to the radial thrust motion for this time interval recorded from kinematic data in the thrust wedge (Platt et al., 1989). It is interesting to note that this radial migration of the thrust wedge–foreland basin system is markedly different from the north-northeastward plate motion of Africa relative to Europe at this time (Dewey et al., 1989), and is likely to be the result of the Adriatic microplate acting independently from Africa and Europe during this time period (Platt et al., 1989).

It is clear from Figure 8 that from late middle to latest Eocene time, the amount of coastal onlap of the Nummulitic Limestones varied along strike. The amount indicated must be considered a minimum value with a 15% error bar based on the problems of section restoration discussed earlier. A mean coastal onlap distance was achieved by taking a value every 20 km along the strike of the stratigraphic onlap from Figure 8. Hence, the minimum value for the amount of mean onlap that occurred during this time interval in a northwest direction from across Switzerland is 46 km, and is likely to be as much as 53 km. From France, the time-equivalent mean onlap amount that occurred in a dominantly westerly direction ranges from 29 to 33 km.

However, error bars must also be incorporated for variations in the time scales used. The upper age bracket is the Eocene-Oligocene boundary, which has been notoriously difficult to date (Prothero, 1990). Harland et al. (1989) dated the Eocene-Oligocene boundary as 35.4 Ma. However, recent ages from a critical section in Wyoming date the boundary as 34–34.5 Ma (Prothero, 1990). The lower age bracket for the onlap information is the boundary between the middle and upper Eocene. This is dated at 39.4 Ma by Haq et al. (1987) and at 38.6 Ma by Harland et al. (1989). By incorporating these two dates for the middle-upper Eocene boundary, and by using the more recent dates outlined by Prothero (1990) for the Eocene-Oligocene boundary, the duration of the upper Eocene ranges from 4.1 to 5.4 m.y.

By dividing the onlap distance by the time taken to cover that distance, rates of onlap can be calculated. For Switzerland, the time-averaged rate of onlap of the cratonic margin sediments during the upper Eocene ranged between 8.5 and 12.9 mm/yr, compared to that of France, which was 4.9 to 8.0 mm/yr.

Width of the Alpine Foreland Basin from Eocene to Miocene Time

A comparison of the thrust front advance rates to the onlap rates through time enables an approximation of variations in the width of the flexural depression. This assumes that the propagation rate of the thrust front was directly linked to the migration of the orogenic margin of the foreland basin; this is a fair assumption if the surface slope angle of the thrust wedge remained constant during this interval. It also ignores the influence of regional or global sea-level changes on basin width. The 40 m.y. duration in the basin evolution that we are studying coincides with a suggested long-term eustatic fall in sea level (Haq et al., 1987). Therefore, it must be recognized that a eustatic fall would counteract the effects of cratonic margin basin subsidence on the coastal onlap record. This would suggest that the cratonic margin of the basin would have migrated even farther away from the orogen than the documented coastal onlap record suggests. However, without detailed knowledge of the basin margin geometry or the amount of sea-level change, this effect cannot be quantified.

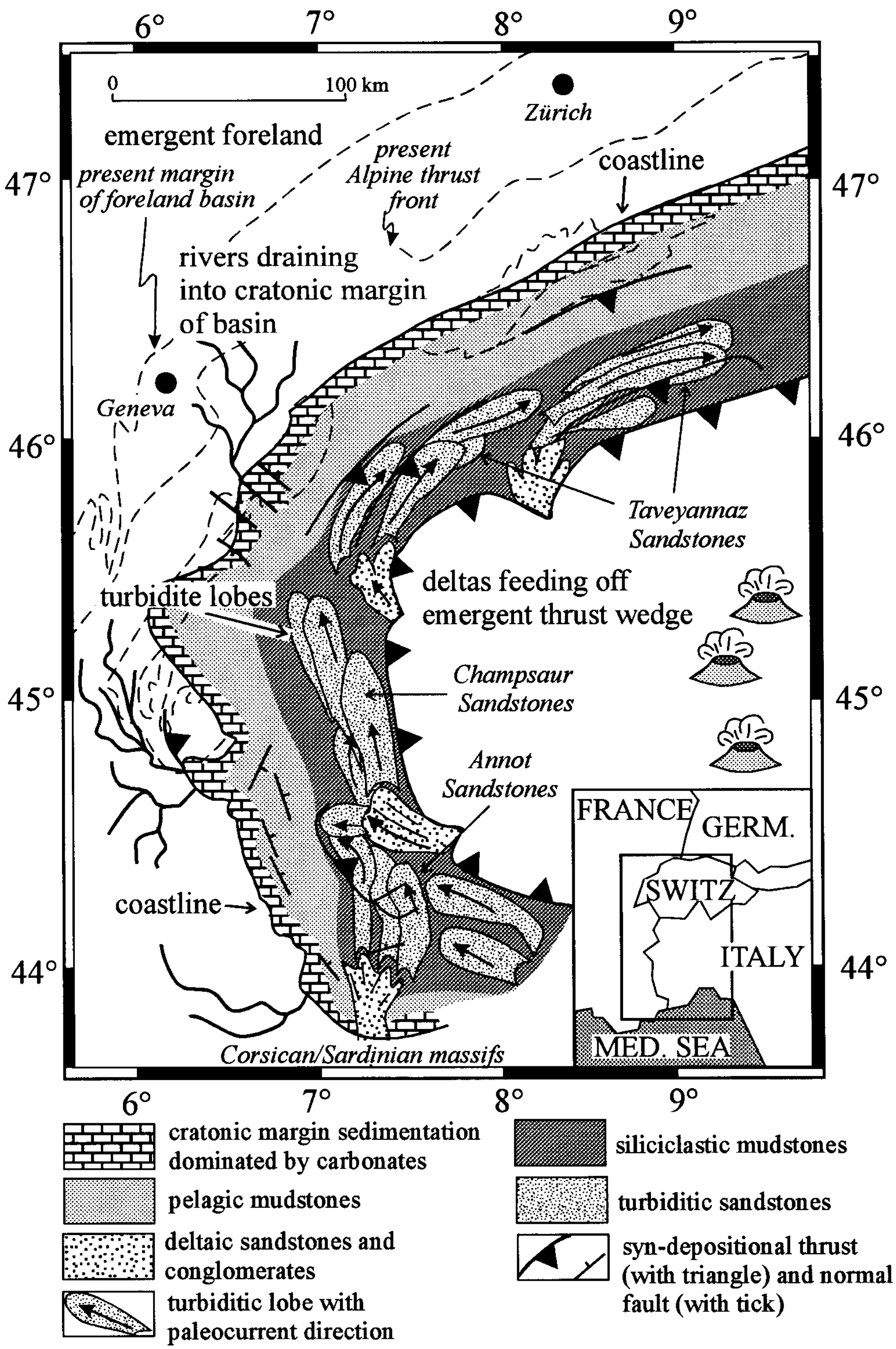

Figure 7. Paleogeographic map of the underfilled Alpine foreland basin during latest Eocene–earliest Oligocene time showing the three main depositional realms. The volcanoes represent the unpreserved island arc inferred from the petrography of the sandstones (De Quervain, 1928; Vuagnat, 1952, 1983; Vernet, 1964; Waibel, 1990).

By plotting the amount of cratonic coastal onlap against thrust front advance for two sections from eastern and western Switzerland (thrust advance data from Sinclair and Allen, 1992, appendices A and B), cross-sectional basin widths are estimated for the full history of the North Alpine foreland basin (Fig. 9). The basin width at any time is taken as the distance from the position of the thrust front to the position of coastal onlap and therefore ignores sedimentation on top of the thrust wedge. The maximum change in basin width of 47 ± 15 to 80 ± 4 km occurred in western Switzerland from late Oligocene to middle Miocene time. However, the long-term trends from the two sections in Switzerland indicate similar rates of thrust

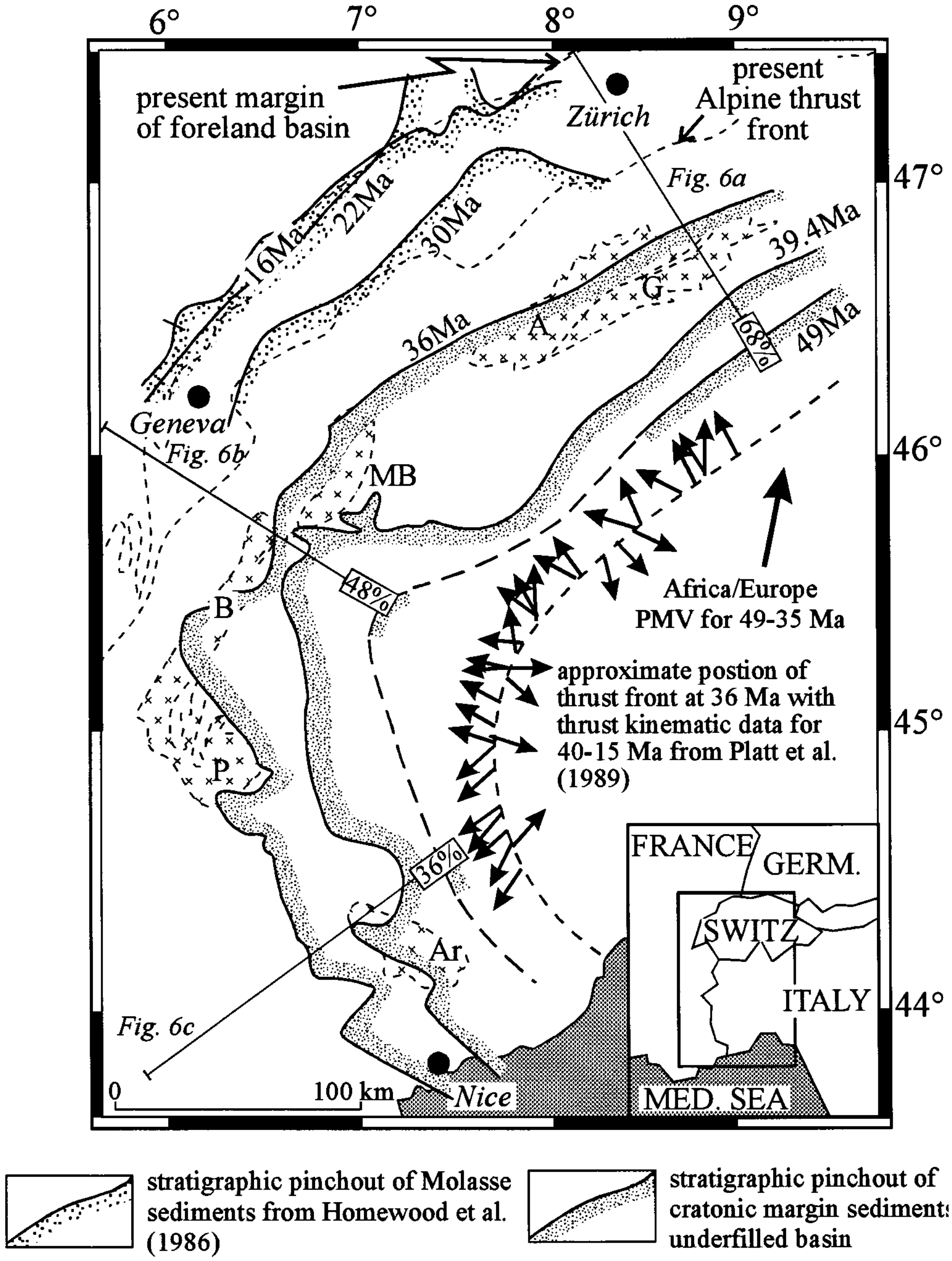

Figure 8. Paleogeographic restoration of the cratonic pinchout of the Alpine foreland basin of France and Switzerland from early Eocene to Miocene time. This uses the carbonates that fringed the early foreland basin and that have been structurally restored (Fig. 6). The Africa-Europe relative plate motion vector is shown for this time interval indicating the variance between it and the radial migration of the thrust wedge–foreland basin system. The boxes at the end of the section lines indicate a minimum value for the percentage shortening of the underfilled foreland basin across that section. Abbreviations as in Figure 4.

front advance and coastal onlap onto the craton. Hence, we can consider the long-term development of the thrust wedge–foreland basin system of Switzerland to have been approximately in steady state (in the sense of Helwig and Hall, 1974, as applied to accretionary wedge-trench systems).

Variations Around the Alpine Arc

Given the arcuate nature of the Alpine chain, it might be predicted that with a north-northeastward relative plate motion of Africa for this time interval (Dewey et al., 1989), the French portions of the basin might reflect more strike-slip behavior. However, the stratigraphic architecture for the underfilled stage involving the superimposition of the three basic units and the migration of the cratonic margin of the basin through time indicate similar basin-forming mechanisms around the arc (Figs. 6, 7, and 8). The results indicate variations in the rates of basin shortening and basin migration that may be more linked to the relative plate motion. In Switzerland, where the collision was orthogonal to the European plate margin, the amount of basin shortening is greater (48%–68%) and the rate of basin migration was faster (8.5–12.9 mm/yr) than in southern France, where basin shortening is 35% and the basin migration rate was 4.9–8.0 mm/yr.

Secondary variations occur with respect to the degree of normal faulting during deposition of the lower and middle units. Some of the best documentation of active normal faulting perpendicular and parallel to the thrust front (Pairis and Pairis, 1975; Lateltin and Muller, 1987) is on the Swiss/

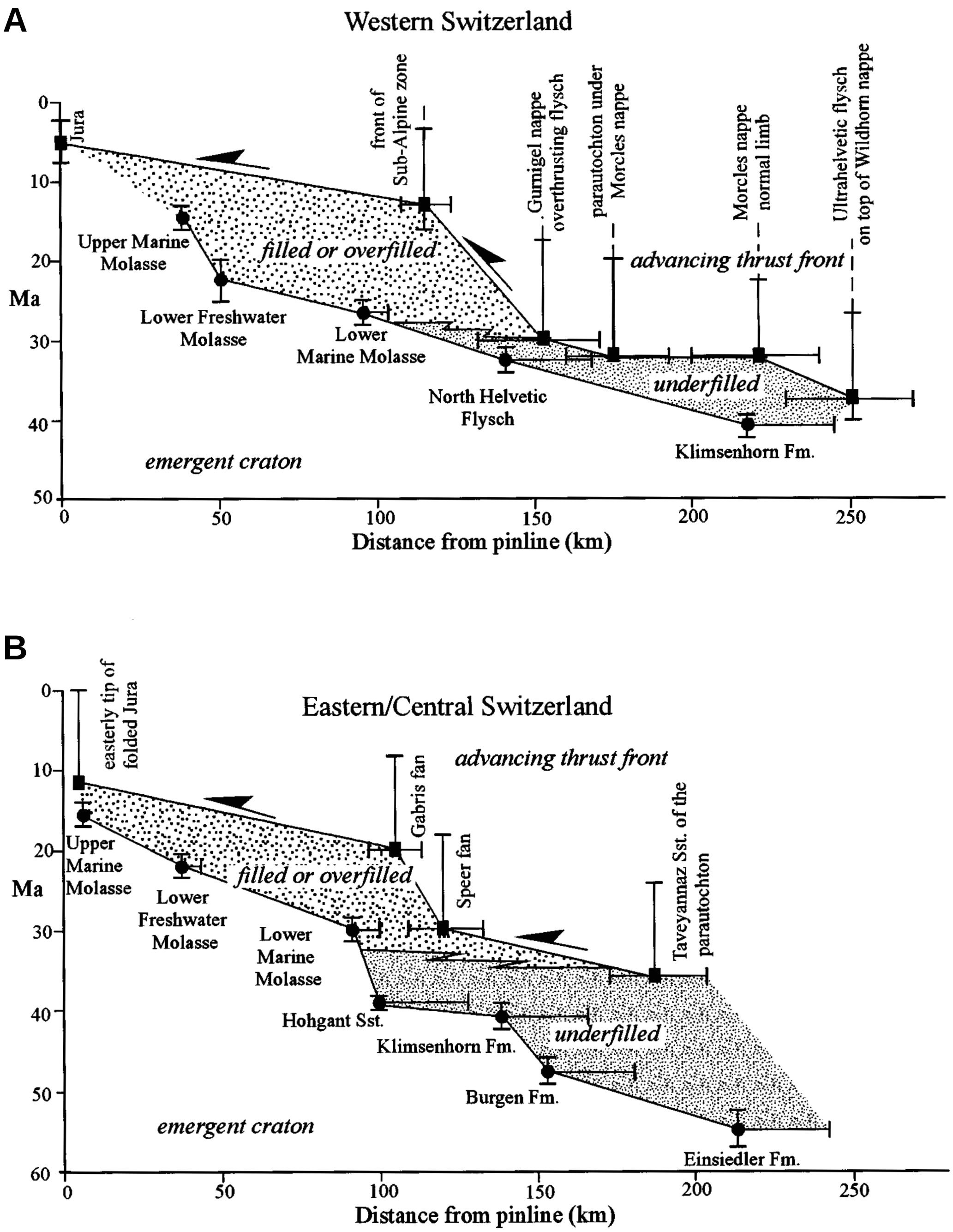

Figure 9. A plot of stratigraphic onlap onto the European craton (from Fig. 6) and thrust front advance (from Pfiffner, 1986; Sinclair and Allen, 1992) against time (Ma) for western (A) and eastern/central (B) Switzerland. Error bars on lower lines for the onlap data include problems of dating and structural restorations. The dating of the thrust advance has broad vertical error bars due to the nature of the dating technique, which involves identifying the youngest age of strata deformed at a point (see Sinclair and Allen, 1992, for data acquisition and evaluation of error bars).

French border and in Haute Savoie. Given that the Rhine-Bresse-Rhône graben system started rifting in the late Eocene (Bergerat et al., 1990), it is possible that these faults are related. The lack of faulting farther east into Switzerland at this time gives support to the influence of rift-related structures in the west.

Farther south in Alpes Maritimes and Haute Provence, the foreland had already been involved in Pyrenean deformation during Late Cretaceous to early Eocene time. These compressional fold structures trend east-west, and are thought to have caused lateral structural variations during subsequent east-west Alpine shortening (Elliott et al., 1985). They continued to influence all three units of the underfilled trinity.

The most marked variation in the structure of the basin around the Alpine arc is from the Pelvoux region, where southwest-directed Alpine thrusting led to the exhumation of cratonic crystalline basement prior to the onset of flexural subsidence (Fig. 7). The tectonic significance of this cratonic thrusting is not fully understood, but it would imply that the earlier development of the basin evolved in a piggyback fashion overlying a deep-seated thrust plane.

CONTROLS ON UNDERFILLED BASIN STRATIGRAPHY

All of the factors that control the development of underfilled peripheral foreland basins also control the later filled or overfilled stage of the basin's development. However, the identification of three distinct depositional realms within an underfilled basin enables the clearer distinction of processes which preferentially influence these different parts of the basin. The controls are separated into regional and local, and are listed below with examples of basins where they have been interpreted to have played an important role.

Regional Controls

(1) Flexural Rigidity of Underlying Plate. The significance of the flexural rigidity of the plate in terms of the broad geometry of foreland basins has been discussed previously (Jordan, 1981; Karner and Watts, 1983). The strength of the European plate underlying the underfilled Alpine foreland basin during the Eocene has been estimated at somewhere below 17 km effective elastic thickness (Sinclair, 1996). The strength of the plate controls the horizontal scale of compensation of the load of the thrust wedge, and hence the width and cross-sectional geometry of the basin (Turcotte and Schubert, 1982). During cratonward migration of the thrust wedge–foreland basin system, the cross-sectional profile of the basement deflection reflects the path of the tectonically induced subsidence of the basin. The rate at which this subsidence occurs for a given flexural rigidity depends upon the rate of horizontal load migration (see below).

For a given flexural profile with a steady load migration rate, the rate of flexural subsidence will increase from the point of zero deflection at the cratonic margin of the plate into the basin as it migrates toward the orogenic margin of the basin (Kominz and Bond, 1986; Allen et al., 1986; Dorobek, 1995). In considering underfilled foreland basins, the acceleration of subsidence leads to a drowning of carbonate platforms of the lower unit on the cratonic margin of foreland basins. The quantitative aspects of carbonate growth versus subsidence rates in foreland basins were described by Dorobek (1995). One of the conclusions of Dorobek's work is that weaker plates should encourage narrower carbonate platforms, which become drowned more rapidly, and that stronger plates generate broader platforms which are able to keep up with subsidence rates for longer periods of time. In the Alpine example the drowning of the Nummulitic Limestones and the generation of water depths greater than 500 m during accumulation of the Globigerina Marls is interpreted to represent the acceleration in flexural subsidence.

Complications in the interaction of the thrust wedge and the foreland plate can arise when the lithosphere is heterogeneous in terms of its strength. Waschbusch and Royden (1992) demonstrated how zones of weakness in the foreland plate cause increased plate curvature and effectively lock the position of the cratonic margin of the basin for greater periods of time than a homogeneous plate model scenario. These workers suggested that in the Apennine and the Kilohigok basins, zones of lithospheric weakness resulted in aggradational facies patterns on the cratonic basin margins.

(2) Thrust Load Migration Rate. The rate at which the distributed load of the thrust wedge propagates over the foreland dictates the rate at which the cratonic margin of the basin subsides; this is a primary control on the degree of filling of the basin (Flemings and Jordan, 1989; Sinclair et al., 1991). In the Alpine example it has been suggested that a decrease in the thrust front advance rates combined with an increase in the exhumation rate of the thrust wedge resulted in the transition from an underfilled to a filled depositional state of the basin (Sinclair and Allen, 1992).

(3) In-Plane Stress. The occurrence of compressional deformation in cratonic interiors some distance away from the zone of flexural deformation and collision is the result of horizontal stresses transmitted through the lithosphere (Letouzey, 1986; Cloetingh, 1988; Karner et al., 1993; Heller et al., 1993). The responses to in-plane forces can be elastic, involving the entire lithosphere leading to modifications in the flexural profile, and/or inelastic (brittle and ductile creep) including the reactivation of faults in the foreland plate. The amplitude and wavelength of in-plane stress-induced deformation is dependent upon the preexisting deflection of the lithosphere. In general, vertical changes in topography of meters to tens of meters is possible, rates of motion being dependent upon the rates of change of the in-plane stress vectors (Heller et al., 1993; Peper et al., 1995).

In considering underfilled foreland basins, modifications of the flexural profile under an imposed compressional horizontal stress would lead to uplift in the region of the peripheral forebulge, and a synchronous subsidence of the basin center (Karner, 1986). The biostratigraphic and paleobathymetric resolution within foreland basin stratigraphy has not enabled the identification of unconformities that can be unequivocally interpreted as responses to in-plane stresses. Therefore, the exact role that in-plane stress has played in controlling the nature of underfilled basin stratigraphy remains unknown.

(4) Eustasy. The interaction of global sea-level change with foreland basin subsidence was outlined by Posamentier and Allen (1993), who separated the basin into zones where basin subsidence would outpace any fall in eustatic sea level, and zones where eustatic fluctuations would dominate over basin subsidence. However, their study lacked any quantitative comparisons of subsidence versus eustasy, and given rates of glacioeustatic induced sea-level change of greater than 45 mm/yr for rises (Blanchon and Shaw, 1995) and up to 5 mm/yr during falls (Williams et al., 1993), these would easily outpace rates of basin subsidence which range from 0.1 to 0.2 mm/yr (Homewood et al., 1986; Cross, 1986). Hence, it should be expected that particularly during periods when ice caps are present, underfilled foreland basins record high-frequency eustatic fluctuations, particularly on their cratonic margins, superimposed on a longer term, subsidence-induced, relative sea-level rise.

The Nummulitic Limestones of eastern Switzerland record a series of transgressive-regressive pulses which have been interpreted as resulting from eustatic fluctuations superimposed on steady flexural basin subsidence (Crampton, 1992; Lihou, 1995). During periods of sea-level highstand, it is possible that the cratonic margin of the foreland basin and the region of the forebulge will be flooded, developing broad, shallow marine conditions. An example of this can be seen in the present Sahul shelf off northwestern Australia, which represents the foreland region to the Timor trough (Audley-Charles, 1986).

In terms of eustatic controls on the orogenic margin of underfilled foreland basins, much can be learned from modern accretionary wedge settings. Stevens and Moore (1985) suggested that the Holocene rise in sea level greatly slowed or even caused the cessation of sediment delivery down submarine canyons of the western Sunda arc accretionary prism to the trench. Therefore, it is feasible that sediment delivery from the orogenic margin of deep marine foreland basins would be strongly influenced by eustasy.

(5) Climate. Sediment production rates in carbonate systems are primarily a function of ocean physiochemistry (Tucker and Wright, 1990). This is illustrated in underfilled foreland basins by the example from Papua New Guinea, where the lower unit was initially dominated by a relatively narrow ramp-type carbonate platform comprising red algal and large foraminiferal grains. Subsequently, with the transition from subtropical to fully tropical conditions, the cratonic margin developed a broad (500 km) rimmed platform (Pigram et al., 1989). In very arid conditions such as during the late Campanian of the southern Persian Gulf, cratonic margin sedimentation comprised a broad platform with a central evaporitic pan (Murris, 1980).

In terms of the orogenic margin of the basin, siliciclastic sediment flux from the mountainous thrust wedge will be influenced by the precipitation amount, the seasonality of the precipitation, vegetation, bedrock lithology, and perhaps most important, the local relief generated by the interaction of climate and uplift (Summerfield and Hulton, 1994).

Local Controls

(6) Cratonic Margin Structure. The inherent structure of the foreland plate plays an important role on a number of scales. As discussed above, on the large scale, the mechanical properties of the lithosphere that may be inherent from previous passive margin processes strongly influence the cross-sectional profile of the basin. Equally, the plan-view geometry and fault distribution play an important role on the cratonic margin subsidence pattern. As described below, normal faults that are oblique to the strike of the Alpine foreland basin in Haute Savoie, France, controlled the distribution of coralgal reef growths of the Nummulitic Limestones (Lateltin and Muller, 1987). Similarly, present-day normal faults with up to 150 m displacement into the axis of the basin are found cutting through the lower unit of calcarenites on the Sahul Shelf south of Timor (Veevers et al., 1978). It has been suggested that active faulting on the foreland plate may be associated with extensional stresses generated on the outer arc of the flexed lithosphere (Bradley and Kidd, 1991).

Uplifted regions on the craton ahead of the foreland basin may provide siliciclastic sedimentation to the cratonic margin and inhibit carbonate growth. The Kilohigok basin of northwest Canada, the Arkoma basin of the Ouachita Mountains, and the Brooks Range foredeep of Alaska are examples of this process. In the Appalachians, the uplift that generated the unconformity overlying the Ordovician Knox and Beekmantown formations was highly variable along strike, and is thought to have been related to reentrants and promontories of the previous passive margin (Lash, 1988).

(7) Orogenic Margin Structure. The distribution and thickness of facies of the upper unit are strongly influenced by the propagation of thrust faults and the generation of lateral structures at the tip of the thrust wedge. Piggyback basins (Ori and Friend, 1984) have been documented as primary controls on the upper unit from the Alps (Apps, 1987; Lateltin, 1988; Sinclair, 1992), the Apennines (Ricci-Lucchi, 1986), and the Pyrenees (Labaume et al., 1985; Mutti, 1985). Lateral structures to thrust faults which underlie piggyback basins may cause lateral compartmentalization of the basin, separating shelf and basin sediments derived from the thrust wedge, as documented from the Pyrenees (Mutti et al., 1988).

ALPINE MODEL FOR UNDERFILLED FORELAND BASINS

By studying the stratigraphic development of the early stages of the Alpine foreland basin, and by integrating this with data from other ancient and modern underfilled basins, it is possible to construct a general qualitative model of the initiation, growth, and demise of an underfilled peripheral foreland basin (Fig. 10). The value of such qualitative models that develop from geologic observations lies in their ability to identify processes that have not yet been fully integrated into the quantitative models, but that are important in terms of the basin's development. The development of the basin has been divided into four stages.

Stage 1. The template upon which peripheral foreland basins develop is the passive margin; in the Alpine case this is represented by the Mesozoic Tethyan succession. Initiation of the foreland basin occurs when load-induced subsidence and sediment derived from the orogenic margin start to influence the outer passive margin setting (Fig. 10, stage 1). At this stage, the encroaching thrust wedge is dominantly deep marine and has many of the characteristics of an oceanic accretionary wedge. Behind the evolving thrust wedge, an active island-arc system may feed volcanic detritus into the trench, as in the present Timor example (Karig et al., 1987), the Appalachian example (Table 1), and the Alpine example during deposition of the Taveyannaz-Annot Sandstones (Fig. 7).

Farther toward the craton, the outer shelf of the passive margin should undergo slow (<0.01 mm/yr) uplift due to the cratonward migration of the forebulge (Crampton and Allen, 1995). Given such slow rates of uplift, wave activity on the shelf can cause abrasion and planation of the topset strata of the passive margin. The deeper water parts of the shelf should undergo shallowing without erosion; however, there has been no evidence for this from the stratigraphic record. Quantitative modeling of forebulge erosion has ignored the process of wave erosion, considering only subaerial processes (Crampton and Allen, 1995). It is clear from the Alpine example that the forebulge region was emergent at some stage as indicated by fluvial valley fills and sidérolithique red beds underlying the Nummulitic Limestones. This may imply that other more rapid mechanisms of relative sea-level change were active at this time on the cratonic margin of the basin (Gupta, 1994).

Stage 2. As the flexural profile migrates over the craton, so the previously uplifted outer shelf of the passive margin begins to subside, and carbonate ramp sedimentation typically takes place (Dorobek, 1995). Depending on the original paleobathymetry and the amount of forebulge uplift, these early carbonates may or may not be separated from the passive margin succession by an unconformity. The zone of forebulge uplift has passed cratonward, uplifting more proximal sediments of the passive margin. The great thickness of carbonates (up to 1200 m) that can accumulate on the cratonic margin of an underfilled foreland basin has yet to be integrated into quantitative models that simulate sedimentation throughout underfilled foreland basins as slope-controlled diffusion of siliciclastic sediment eroded from the thrust wedge (Flemings and Jordan, 1989; Sinclair et al., 1991; Jordan and Flemings, 1991; Johnson and Beaumont, 1995).

Offshore from the carbonate ramps, hemipelagic muds are deposited (middle unit). This reflects the location at which flexural subsidence exceeds the growth rate of the carbonate platform. Dorobek (1995) estimated that a subsidence rate of 1 mm/yr over a time period of 10^3–10^5 yr would lead to drowning of most carbonate platforms, and that the transition from shallow water benthic carbonate production to pelagic-hemipelagic mudstones would occur in water depths of ~100 m (Hallock and Schlager, 1986). This region basinward of the neutral point of the flexural profile undergoes maximum outer arc flexural extension and hence may develop or, more likely, reactivate normal faults, as documented in the Taconic foreland of New York (Bradley and Kidd, 1991), the Arkoma basin (Houseknecht, 1986), the Timor trough (Veevers et al., 1978), and the Alps (Lateltin and Müller, 1988; Elliott et al., 1985).

Stage 3. Continued convergence and migration of the flexural profile leads to the migration of the three depositional realms across the craton, resulting in the superposition of the underfilled trinity. The carbonate platform onlaps the cratonic margin at a comparable rate to the migration of the thrust front (Fig. 9) indicating a steady-state wedge-basin system (see Alpine example). Stratigraphic cyclicity within the carbonates are generated by high-frequency sea-level fluctuations which may be eustatic or possibly associated with in-plane stress fluctuations.

Early deposits on the orogenic margin of the basin are incorporated into the thrust wedge by the process of frontal accretion. With increased exhumation rates, the sediment supply from the thrust wedge increases, and deltas prograde away from the thrust wedge and into the basin.

Stage 4. A change in the dynamics of the thrust wedge-foreland basin system transforms it from an underfilled state to one in which sediment flux outpaces the generation of accommodation space, so filling the basin. This is usually achieved by filling of the basin along its axis, following the regional depositional gradient as recorded in the Alaskan, Alpine, and Pyrenean examples. In the Alpine example it is believed that a change in

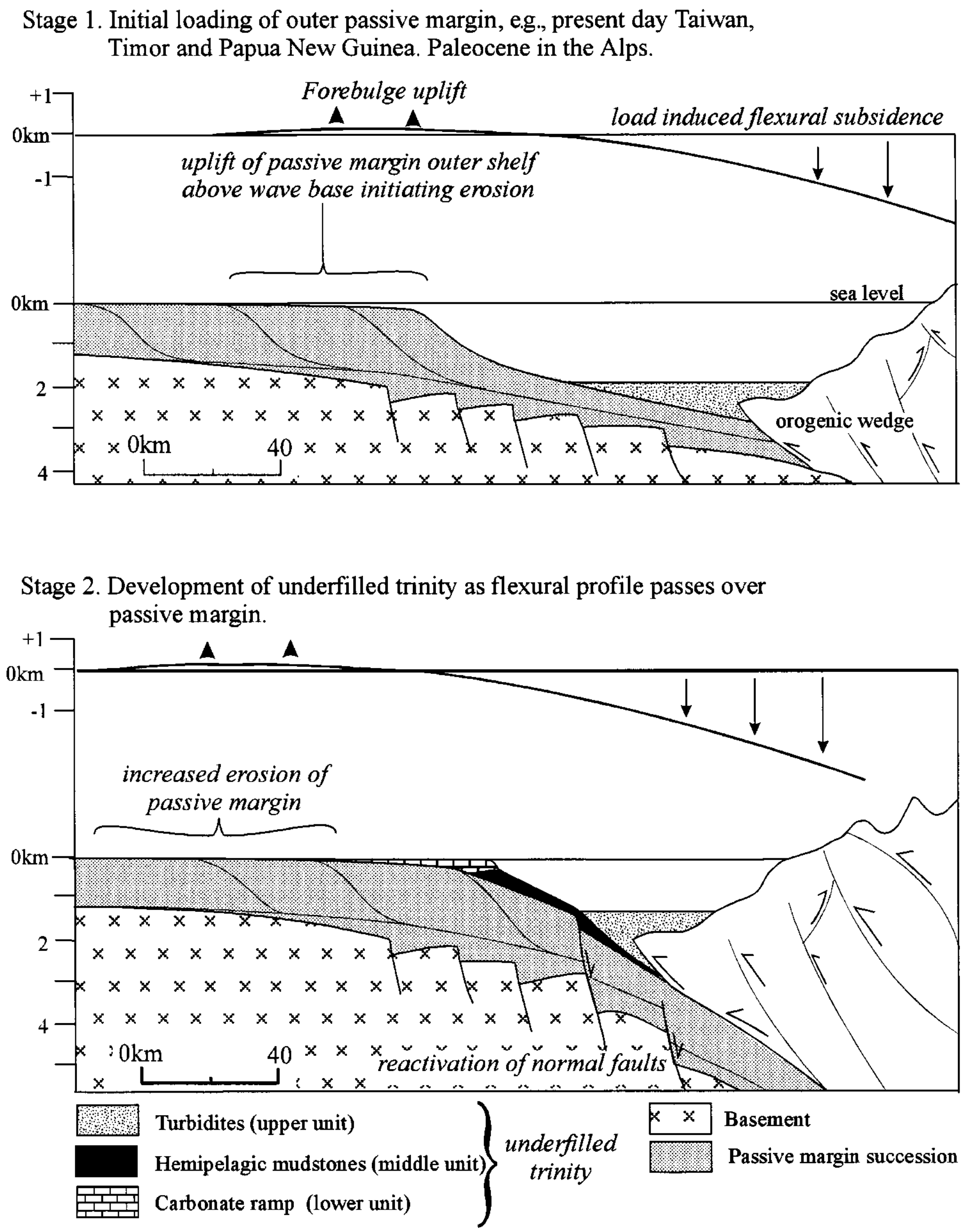

Figure 10. A general evolutionary model for the tectonic and stratigraphic development of underfilled peripheral foreland basins based primarily on the Alpine example. See text for description of the four stages.

the thrust wedge dynamics associated with increased backthrusting and decreased frontal advance leading to higher exhumation rates caused the filling of the basin (Sinclair and Allen, 1992). However, other mechanisms such as increased flexural rigidity of the plate (Watts, 1992) or the filling of the inherited passive margin bathymetry (Stockmal and Beaumont, 1987) may be applicable to other basins.

CONCLUSIONS

(1) The stratigraphy of underfilled peripheral foreland basins can be synthesized into three basic units, which are commonly superimposed during basin migration. The three units are here termed the "underfilled trinity"; the lower unit reflects the accumulation of carbonates on the cratonic margin of the basin, the middle unit reflects hemipelagic fall-out of muds offshore from the cratonic margin, and the upper unit reflects turbiditic siliciclastic sedimentation on the cratonic margin of the basin, classically termed flysch. However, variations occur from the above simplification, particularly in the lower unit. In the Eocene of the underfilled Alpine foreland basin, the three units of the trinity are superimposed on top of one another and are clearly distinguishable within the deformed northern and western margins of the orogen.

(2) The degree of structural shortening of the Alpine underfilled foreland basin decreases westward and southward around the Alpine arc; maximum values are ~68% in eastern Switzerland, 48% in Haute Savoie, eastern France, and 35% in Haute Provence, southeastern France.

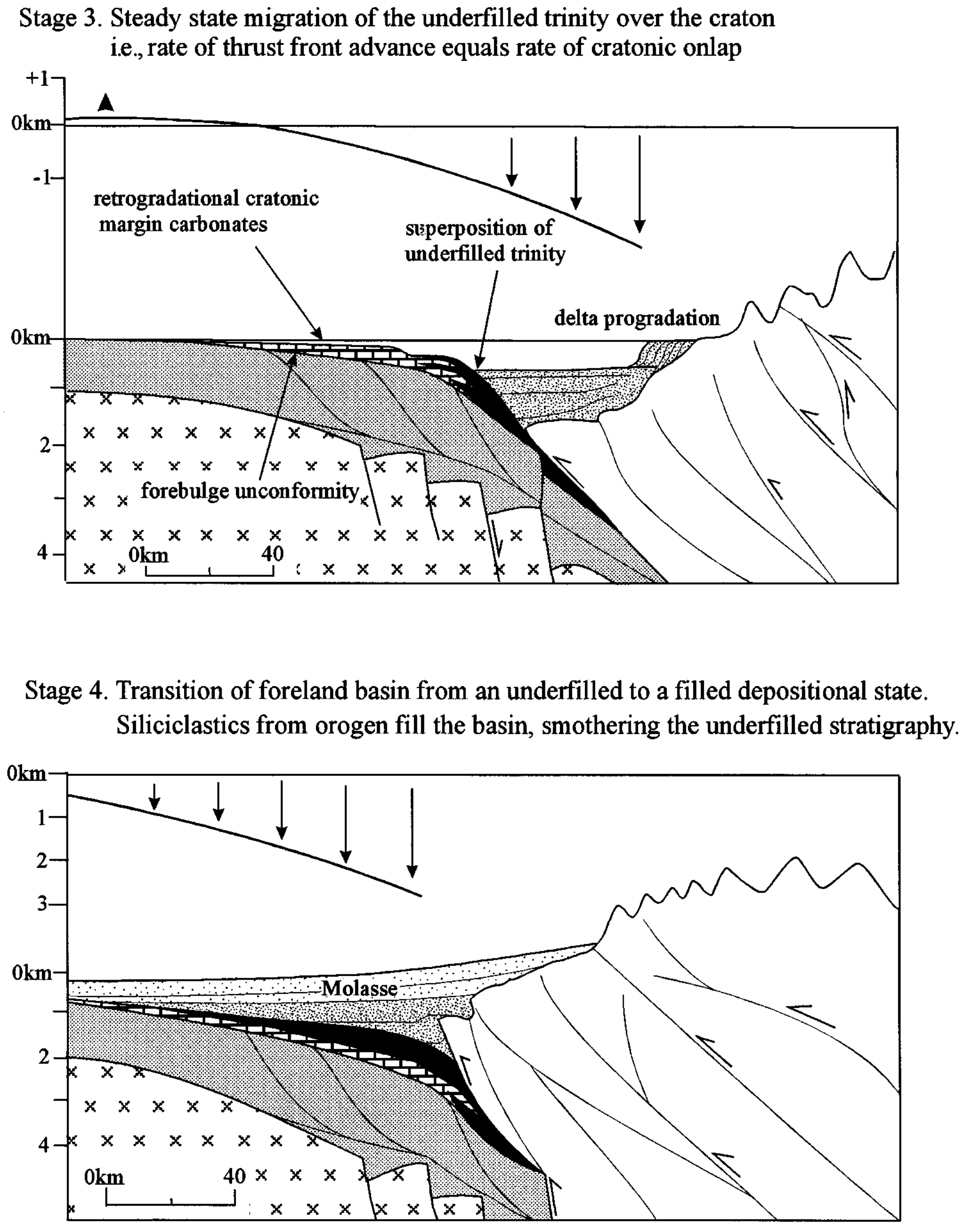

Figure 10. (*Continued*).

(3) The paleogeographic restoration of the Alpine basin allows coastal onlap rates of the lower unit (Nummulitic Limestones) onto the European craton to be calculated. In broad terms, during late Eocene times, the time-averaged rate of north to northwestward coastal onlap in Switzerland was between 8.5 and 12.9 mm/yr. The time-equivalent westward to southwestward onlap rate from France was between 4.9 and 8.0 mm/yr. The directions of basin migration over the craton were parallel to the time-equivalent thrust motions measured from within the orogen (Platt et al., 1989), but were oblique to the Africa-Europe plate motion vector (Dewey et al., 1989).

(4) The rate of coastal onlap of the cratonic margin of the underfilled basin in Switzerland can be combined with stratigraphic pinch-out migration rates during the overfilled history of the basin (Homewood et al., 1986). This record of motion of the cratonic edge of the basin is compared to the rate of thrust propagation into the orogenic margin of the basin (data from Sinclair and Allen, 1992). The two rates are broadly comparable, indicating that the Alpine foreland basin of central Switzerland migrated in a steady state for a distance of ~210 km over the European craton from early Eocene to middle Miocene time. The basin of western Switzerland migrated ~170 km from late Eocene to middle Miocene time.

(5) Variations in the development of the foreland basin around the Alpine arc include decreased rates of basin migration and amounts of shortening from east to west and south, and the increase in syndepositional fault activity toward the west and south. The latter is thought to have been linked to the synchronous opening of the Rhine-Bresse-Rhône graben system.

(6) By combining the Alpine study with other studies of ancient and modern underfilled foreland basins it is possible to construct a general evolutionary model for the growth of a typical underfilled foreland basin and its subsequent filling. This model describes the variations in depositional surface processes that occur in underfilled basins, but that have not yet been fully integrated into recent numerical models.

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