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Basement-involved shortening and deep detachment tectonics in forelands of orogens: Insights from recent collision belts (Taiwan, Western Alps, Pyrenees)

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[1] New combined structural and seismotectonic analyses demonstrate basement-involved shortening in forelands of recent collisional orogens (Taiwan, Western Alps, Pyrenees). Basement thrusts documented by seismicity (e.g., the 21 September 1999, Chi-Chi earthquake in Taiwan) and/or structural data are triggered and localized by preexisting basement faults which constitute crustal weakness zones available for reactivation under low stress levels. Reactivation of basement faults may induce localization of folds and thrusts in the shallow thrust wedge, development of crystalline thrust sheets, out-of-sequence basement thrusting and late basement uplift, deformation/refolding of shallow thin-skinned nappes, and development of accommodation structures such as transfer faults leading to a kinematic segmentation of foreland thrust belts. Reactivation of preexisting basement faults also occurs in the far foreland in response to the far-field transmission of orogenic stresses depending on the amount of the changing-through-time mechanical coupling between the orogen and its foreland. Displacements related to basement shortening in forelands are accommodated at the scale of the upper crust, which requires that it is partially decoupled from the deeper lithospheric levels by a crustal detachment. This detachment presumably occurs along the midcrustal, thermally weakened brittle-ductile transition. It may either ramp toward the surface into a shallow, upper crustal detachment beneath the fold-thrust belt and/or extend beneath the foreland and accommodate basin inversion far away from the orogen. Occurrence and relative timing of shallow and deep detachment tectonics in forelands seem to be dependent on mechanical boundary conditions such as the presence of ductile horizons within the cover sequence or of preexisting weakness zones in the underlying basement. *INDEX TERMS:* 7230 Seismology: Seismicity and seismotectonics; 8102 Tectonophysics: Continental contractional orogenic belts; 8164 Tectonophysics: Stresses—crust and lithosphere; 8010 Structural Geology: Fractures and faults; 8120 Tectonophysics: Dynamics of lithosphere and mantle—general; *KEYWORDS:* seismotectonics, thick-skinned tectonics, forelands, far-field orogenic stresses, Taiwan, Western Alps

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1. Introduction and Scope of the Study

[2] The overall geometry and mechanics of foreland thrust belts are well accounted for by the Coulomb critical wedge model [*Davis et al.*, 1983; *Dahlen et al.*, 1984]. This model describes these fold-thrust belts as critically tapered wedges made of homogeneous deformable material moving above an undeformed substratum along a shallow, low dipping detachment (Figure 1a). *Davis et al.'*s [1983] model, however, fundamentally meets three main restrictions. The first one deals with the debatable mechanical analogy between foreland thrust belts and oceanic accretionary sedimentary wedges, as emphasized by *Bombolakis* [1994]. The second one is related to the assumed homogeneous nature of the deformable material of the wedge: This concept can be generalized to a material with distributed preexisting weaknesses so their reactivation may result in a distributed strain resembling at a large scale that of an homogeneous material, but fails in the case of the reactivation of few localized inherited fault zones. The third one lies in the assumed rigid and undeformable behavior of the substratum below the ductile basal horizon: The deformable material is often thought to be restricted to the detached cover, which leads to favoring thin-skinned tectonics and to often underestimating basement-involved shortening in forelands of orogens.

[3] A number of regional studies have demonstrated the compressional reactivation of preexisting structures within both the cover and the basement of foreland thrust belts (e.g., Alps [*Roure et al.*, 1990]; Urals [*Brown et al.*, 1999]; Andes [*Winslow*, 1981; *Kley et al.*, 1999; *Cristallini and Ramos*, 2000]; Zagros [*Jackson*, 1980; *Berberian*, 1995]; Rockies [*Dechesne and Mountjoy*, 1992]; Taiwan [*Mouthereau et al.*, 2002; see also *Letouzey*, 1990; *Mitra and Mount*, 1998]). This raises the question of the way thin-skinned and thick-skinned related structures interact in space and time. Moreover, within-plate reactivation of basement faults and structural inversion of sedimentary basins have also been documented in the foreland far from the orogens [e.g., *Tapponnier et al.*, 1986; *Ziegler*, 1987; *Ziegler et al.*, 1995; *Lacombe and Mouthereau*, 1999; *Marschak et al.*, 2000].

[4] The two-fold aim of this paper is (1) to provide further evidence of basement-involved shortening in foreland thrust belts of recent collisional orogens and (2) to address the problems of the way displacements related to

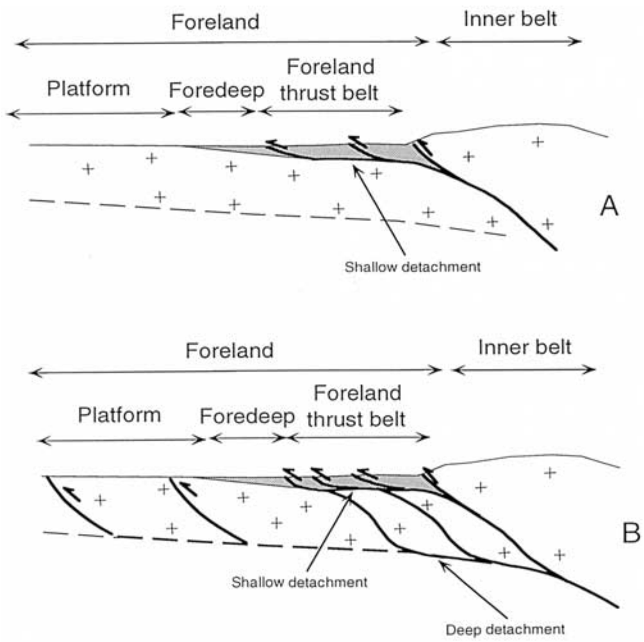


Figure 1. Conceptual view of tectonic styles in the forelands of orogens. (a) Shallow detachment tectonics. The fold-thrust belt mainly consists of the shortened cover detached from the nearly undeformed basement above a shallow basal ductile horizon. (b) Shallow and deep detachment tectonics. The fold-thrust belt is made of both cover units detached above a shallow basal ductile horizon and basement thrust units detached above a low-friction deep crustal detachment. Both the cover and the basement are involved in shortening, but display different styles of deformation.

structural inversion/basement fault reactivation can be accommodated and orogenic stresses transmitted in the foreland far away from the orogens through deep crustal detachment. Attention will be paid to the timing of basement fault reactivation and coeval activation of deep detachments and to the links between compressional foreland deformations and mechanical coupling between the orogen and its foreland. For this purpose, we study three examples of forelands of orogens that differ on several aspects but mainly on the age of tectonism: They are the western Foothills of the Plio-Quaternary and still active Taiwan collision, the western Alpine foreland thrust belt which formed mainly during the Mio-Pliocene and display evidence of still active deep deformation, and the northern foreland thrust belts of the Pyrenees that formed during the late Eocene. Because of the difference in age of tectonism, these mountain belts presumably provide different signatures of basement-involved deformation and therefore expectedly reflect different stages or at least various aspects of the structural evolution of forelands.

2. Signature of Basement-Involved Shortening in Foreland Thrust Belts

[5] Reactivation of basement faults occurs during orogenic evolution of collided passive margins, and this structural process is known to exert a strong control on the evolution of orogens [*Dewey et al.*, 1986]. Basement fault reactivation may induce localization of thrusts and folds in the developing shallow thrust wedge, reversal of extensional faults and development of crystalline thrust sheets, out-of-sequence thrusting and refolding of shallow nappes, development of accommodation structures such as lateral ramps, and development of basement uplifts [*Wiltschko and Eastman*, 1983; *Glen*, 1985; *Cooper and Williams*, 1989; *Roure et al.*, 1990; *Narr and Suppe*, 1994; *Butler et al.*, 1997].

[6] In foreland thrust belts of young but no longer active orogens (e.g., Pyrenees-Provence), these signatures can be identified in some places by careful structural investigations of the relationships between cover and basement. In more recent orogens (Western Alps), active basement uplifts recognized by geodetic or seismotectonic investigations may complement structural analyses in demonstrating deep basement thrusting. In still active orogens (Taiwan), seismicity combined with structural analyses provides first-order constraints on deep crustal deformation. In all cases, the study of reversal of preexisting basement (normal) faults is generally much easier in forelands than in inner parts of orogens where the initial relationships between the basement and its sedimentary cover have generally not been preserved and the initial attitude of the faults has been strongly modified or erased by later evolution.

3. Basement Thrusting and Deep Detachment Tectonics in the Active Foreland Thrust Belt of Western Taiwan

3.1. Structural Evidence

3.1.1. Structural Inversion in the Taiwan Foreland

[7] The active fold-thrust belt of western Taiwan results from the oblique collision of the Luzon arc with the Chinese passive margin since 4–5 Ma [e.g., *Suppe*, 1981] (Figure 2a). Beneath the Pliocene-Pleistocene synorogenic deposits of the western foreland basin, the collided Chinese margin displays series of ENE oriented Oligo-Miocene horsts and basins inherited from the opening of the South China Sea (Figure 2b). The precollisional Taihsi and Tainan basins, filled with clastic sediments of the Chinese passive margin, alternate with the Kuanyin and Peikang Highs which represent shallow parts of the pre-Neogene Chinese continental basement (more precisely of the pre-Neogene preorogenic substratum which is, if not undoubtedly the true crystalline basement, a good proxy for it) [*Mouthereau et al.*, 2002].

[8] The influence of the basement on the geometry and kinematics of the orogenic wedge is emphasized by the sigmoidal shape of the belt front (Figures 2a and 2b), which reflects along-strike variations of the thrust wedge geometry and of the structural style of the frontal units [*Lu et al.*, 1998; *Mouthereau et al.*, 2002]. Sandbox models of oblique collision of a deformable medium by an asymmetric indenter [*Lu and Malavieille*, 1994] match the overall shape of the surface traces of structures within the Taiwan thrust wedge, but fail to reproduce the sigmoidal shape of the frontal units and especially the arcuate

shape of the NW Foothills. This strongly suggests that the structure of the western Taiwan Foothills results, in fact, from the superimposition of the oblique impingement of the wedge by an asymmetric hinterland indenter (the accreted part of the Luzon arc–the Coastal Range and (?) the Central Range) and its development onto the irregular shaped Chinese margin made of preorogenic basins and basement highs.

[9] Recent structural investigations [*Hung et al.*, 1999; *Mouthereau et al.*, 2001a, 2002] in the western Taiwan Foothills have demonstrated that reversal of basement normal faults inherited from the preorogenic passive margin history occurred, not only in the inner domains where it is deduced from accurate analysis of outcrop data, but also in the outermost thrust units and even beneath the undeformed Coastal Plain, as evidenced by seismic data. Structural inversion is widely expressed on the southern edges of the Kuanyin and Peikang basement highs (Figure 2b). The cross section of Figure 2c in the Chiayi area, i.e., in front of the Peikang High, was constructed on the basis of wells and available seismic data [*Chang et al.*, 1982; *Chang et al.*, 1996]. It highlights structural inversion beneath the Coastal Plain in the Tainan basin and the possible development of basement thrust sheets beneath shallower, thin-skinned nappes in the western Foothills. A broad foreland syncline is observed, and its eastern limb is affected by back thrusting, which supports occurrence of a deep triangle zone resulting from basement imbricates and deep thrusts which probably root into a 10–12 km deep detachment surface (Figure 2c). The Meilin thrust marks the front of the orogenic wedge. The deformation mode of the Foothills is consequently characterized by superimposed shallow and deep detachment tectonics.

3.1.2. Transfer Faulting and Along-Strike Segmentation of the Taiwan Foreland

[10] A striking signature of the involvement of the basement in collisional shortening is the structural and kinematic segmentation of the Taiwan belt, underlined by the sigmoidal shape of the thrust belt front. Major fault zones, oblique to the structural grain of the belt, recognized as transfer fault zones by *Deffontaines et al.* [1997], play an important role in this segmentation (Figure 2b).

[11] Transfer (or tear) faults are common tectonic features in foreland thrust belts. A thrust sheet moving over a detachment level may exhibit development of transverse structures in response to a lateral change in cover thickness because of basement geometry or basin boundary, to the lateral termination of the main detachment surface which can be viewed as a high/low friction boundary, or to the reactivation of an underlying basement fault. A primary tear fault may thus appear contemporaneously with fold-thrust development, and associated with a curvature of these structures as soon as they form. On both sides, the amount of shortening is the same, but it may be accommodated in a different way. Such a system may secondarily evolve gradually into a transfer (or tear) fault, associated with a clear offset of already formed or currently forming folds, and on both sides of which differential displacement/shortening may be accommodated.

[12] The Chishan Transfer fault zone, a major transverse feature in the SW Taiwan Foothills (Figure 2b), is a good example of such transfer zones. The tectonic evolution and significance of this fault zone have been recently discussed [*Lacombe et al.*, 1999]. It is probably inherited from the preorogenic rifting period and corresponds to an important lateral change of basement depth and thickness of Neogene sediments. Structural evidence alone cannot unambiguously demonstrate that the basement fault zone was reactivated at depth, although this reactivation is supported by seismicity (see section 3.2.2). However, the associated basement top offset has localized in the cover a primary tear fault which accompanied thrust emplacement. This zone secondarily evolved into a transfer fault zone with a dominant left-lateral component of motion, on both sides of which largely contrasting styles of deformation occur [*Mouthereau et al.*, 2001a]. At a more regional scale, this fault zone constitutes a major boundary between a northern domain of actual crustal collision where basement-involved shortening and structural inversion occur and a southern domain which can be regarded as the northern extension of the Manila accretionary wedge and where deformation within the outer units occurs above a low dipping and shallow detachment surface above the pre-Miocene basement [*Lacombe et al.*, 1999].

[13] Another aspect of the segmentation of the belt in response to the reactivation of basement faults can be documented in the NW Taiwan Foothills north of the Sanyi-Puli Transfer fault zone (Figure 2b), where low-angle thrusting interacts with high-angle thrusting related to reversal of basement normal faults inherited from the extensional history of the margin. The NW Taiwan belt consists of folds and west verging thrust sheets composed of precollisional Miocene deposits overlain by synorogenic Plio-Pleistocene formations. The belt front trend swings a full 90° angle from a N160° direction to the south to a N070° direction to the north, and forms a salient (Figure 3a). This segment illustrates nicely the control exerted by the basement on the geometry and kinematics of the foreland thrust belt: not only the salient formed in response to the along-strike variation in the preorogenic basin thickness, and developed in the portion of the foreland basin that was initially thicker (the Taihsi basin), but additionally curvature was largely controlled by the transpressional reactivation of extensional structures along the southern edge of a major basement high of the margin, the Kuanyin High.

[14] In detail, field studies and seismic reflection profiling reveal two main structural trends. In the offshore Taihsi basin, most features strike ENE-WSW [*Huang et al.*, 1993], at high angle to the structural grain of the belt. They correspond either to Paleogene to Miocene normal faults which probably extend down to the Mesozoic crystalline basement, or to few high-angle thrusts which originated from the transpressional reactivation of the previous normal faults [*Yang et al.*, 1994, 1997]. These faults extend onshore in the outer Foothills as parallel high-angle thrusts, along which oblique folds indicate a significant amount of right-lateral wrench movement. Their close association with north and south dipping normal faults, as well as the detailed analysis of thickness and structural elevation of Miocene

strata on both sides of the faults [*Yang et al.*, 1996] demonstrate that these high-angle thrusts result from the compressional reactivation of preexisting normal faults of the margin [e.g., *Suppe*, 1984]. In the inner part of the Foothills, these high-angle faults connect and/or intersect a system of NNE folds and low-angle thrusts parallel to the general trend of the belt (Figure 3a), as documented by seismic lines (Figure 3b). Such nucleation of high-angle thrusts at the upper part of reactivated normal faults and/or connection of high-angle thrusts with low-angle thrusts (Hsinchu fault; Figure 3b) are, together with basement short cuts, common features in basement-involved compressional tectonics.

[15] Stress trajectories in the shallow thrust sheets have been determined from paleostress reconstructions based on tectonic analysis of minor fault sets (Figure 3C). Although the regional transport direction inferred from both the N020° trending structural grain of the belt (major fold axes and thrust traces) and from GPS data [*Yu et al.*, 1997] remained nearly constant (N110°–120°), shortening directions evolved with time in the northern part of the arc from a N120° to a N-S trend, but remained constant in its southern part [*Lacombe*, 2000]. The divergence between shortening and regional transport in the northern part of the arc supports significant wrench deformation parallel to both the arc limb and the structures of the margin (Figure 3a).

[16] The tectonic evolution of the NW Taiwan belt and the control by the reactivated basement normal faults can be summarized as follows: Some of the N070° to E-W normal faults of the Taihsi basin became lateral ramp surfaces of low-angle thrust faults and/or were obliquely inverted and become high-angle thrusts with a right-lateral wrench component. The transpressional reactivation of normal faults in the outer belt predated the activation of shallow detachments and related low-angle thrusting in the inner belt. Thin-skinned thrust sheets that formed in relation to shallow detachments (the lower Miocene Peiliao formation or the lower Pliocene Kueichulin formation [*Namson*, 1984; *Yang et al.*, 1996]) propagated laterally and obliquely toward the reactivated normal faults while deforming internally under a nearly uniform N100°–120° shortening (stage 1, Figure 3c). In a second stage (stage 2), the reversed normal faults were used as lateral/oblique ramps for the low-angle thrusts which continued to propagate; the trend of low-angle thrusts therefore evolved northward into a direction parallel to the strike of the high-angle thrusts, thus generating curvature. This stage 2 was associated with a divergence of the shortening trajectories from the regional transport direction, revealing transpressional deformation along the northern limb of the arc related to the development of a regional-scale oblique ramp (Figure 3c). Displacements related to the motion along the wrench/high-angle thrusts extending down to the pre-Tertiary formations and still active today as deduced from geomorphology [e.g., *Ku*, 1963; *Tang and Hsu*, 1970], although of a maximum several kilometers, were presumably accommodated within the upper crust above a deep, still active detachment at 10–12 km depth (Figure 3b), as suggested by the distribution with depth of microearthquakes [*Lin et al.*, 1989] (Figure 4).

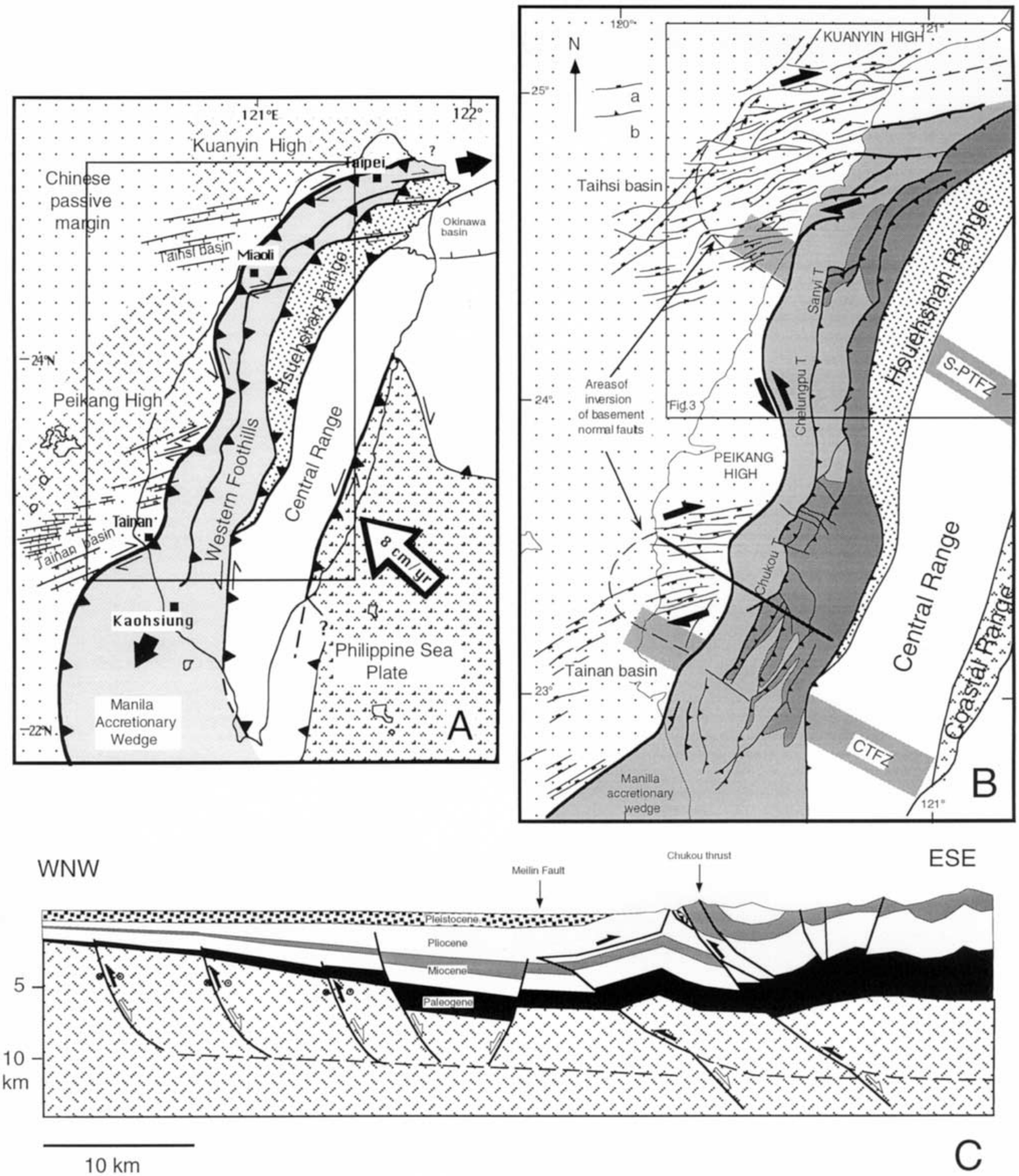
3.2. Seismotectonic Evidence

3.2.1. Seismotectonic Signature of Basement-Involved Shortening

[17] In the Taiwan foreland thrust belt, the ongoing collisional deformation is marked by a high seismic activity (Figure 5). The accurate analysis of the distribution of earthquakes and microearthquake over the period September 1999–February 2000 (data from Central Weather Bureau, Taiwan) provides new insights into the mechanical behavior of the deforming crust in the foreland of the Taiwan orogen, and supports basement-involved shortening and deep-seated detachment tectonics previously discussed based on structural analyses.

[18] In the south central western Foothills, one of the main structural features is the active Chukou thrust, the major crustal boundary between the Foothills and the Coastal Plain (Figures 5 and 6a). East of the Chukou thrust, seismicity shows a nearly homogeneous distribution from the surface up to 15 km depth, with sometimes high-magnitude earthquakes, which indicates pervasive crustal deformation in the Foothills. West of the Chukou thrust, the 5–6 km thick sedimentary foreland deposits of the Coastal Plain are nearly aseismic; below, earthquakes occur up to 20 km depth. Even though their magnitudes are lower than in the Foothills, they show that the basement is actively deforming. This basement seismicity reflects the present-day activity of high-angle thrusts with a right-lateral component of motion ("b" focal mechanisms of Figure 5a), therefore supporting structural inversion on the southern

Figure 2. (opposite) (a) Tectonic setting and main structural features of Taiwan. The large open arrow shows the direction of convergence of the Philippine Sea plate relative to the Chinese passive margin. Bold lines indicate major thrusts, triangles on upthrown side; lines with double arrows indicate wrench faults. The shaded pattern represents the Foothills. Hatched areas correspond to basement highs of the Chinese margin underlying the deposits of the foreland basin. Solid arrows indicate tectonic escape at the tips of the belt. The frame shows the location of Figure 2b. (b) Simplified geological map of the western Foothills. Note the sigmoidal shape of the thrust front (bold line) and the occurrence of inversion tectonics on the southern edges of the Kuanyin and Peikang basement highs. Abbreviations are as follows: S-PTFZ, Sanyi-Puli Transfer fault zone; CTFZ, Chishan Transfer fault zone. Solid arrows underline the areas of major wrench tectonics. Lines corresponding to label "a" indicate normal fault inherited from the passive margin history; lines corresponding to label "b" indicate reactivated normal fault. (c) Geological cross section in the Chiayi area (location on Figure 2b). Note the transpressional reactivation of normal faults beneath the Coastal Plain in the Tainan basin.



121°E
122°
Kuanyin High
Taipei
Chinese passive margin
Okinawa basin
Taihsi basin
Miaoli
Hsuehshan Range
24°N
Peikang High
Western Foothills
Central Range
Tainan
Tainan basin
8 cm/yr
Kaohsiung
Philippine Sea Plate
Manila Accretionary Wedge
22°N
A
120°
N
KUANYIN HIGH
21°
25°
a
b
Taihsi basin
Sanyi T
Hsuehshan Range
Chelungpu T
S-PTFZ
Areas of inversion of basement normal faults
24°
Fig.3
PEIKANG HIGH
Chukou T
Central Range
Tainan basin
23°
CTFZ
Coastal Range
Manilla accretionary wedge
121°
B
WNW
Meilin Fault
Chukou thrust
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Pleistocene
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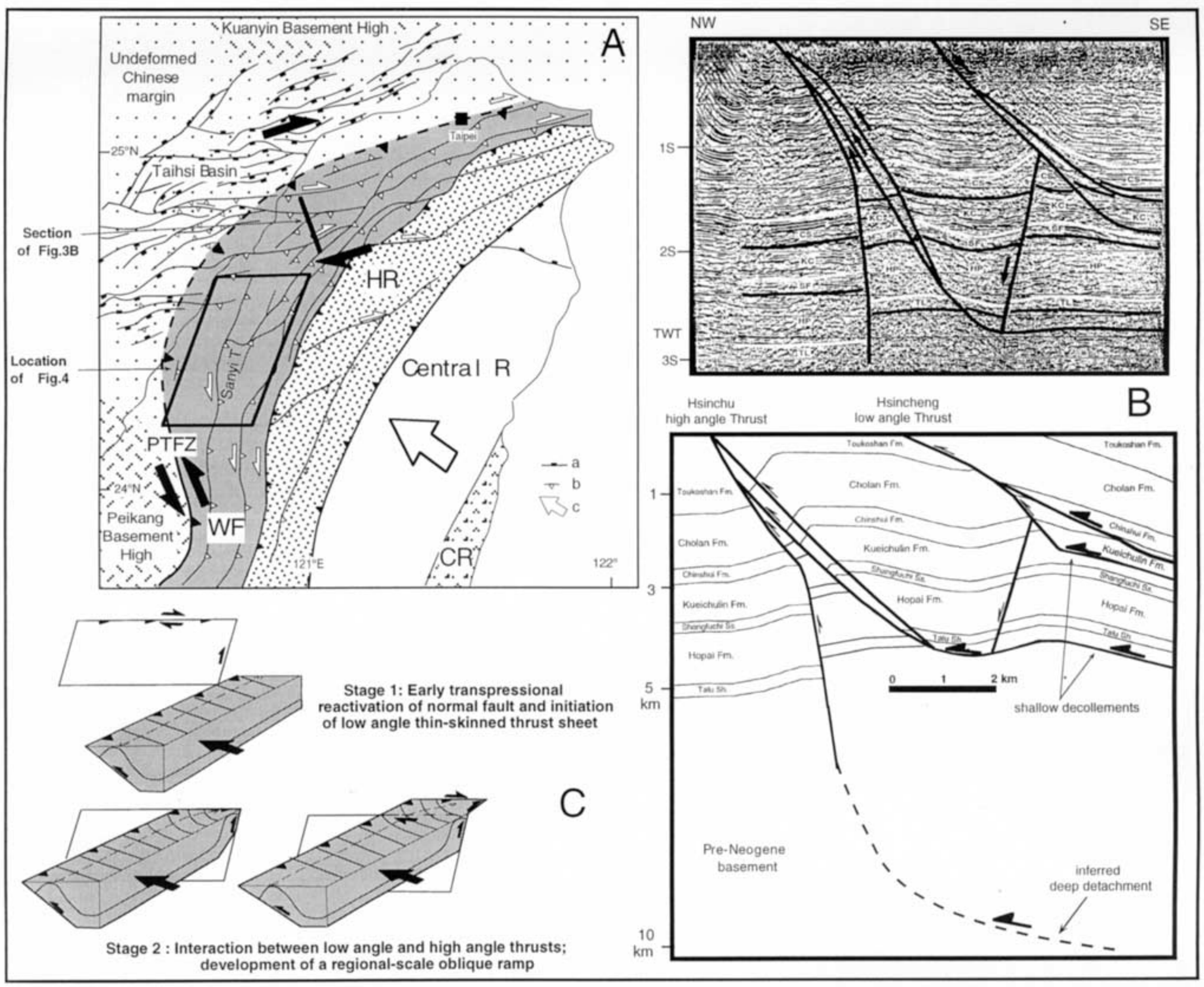


Figure 3. (a) Simplified geological map of the NW Taiwan arcuate belt (see location on Figure 2b. Lines corresponding to label "a" indicate normal fault inherited from the passive margin history; lines corresponding to label "b" indicate reactivated normal fault; open arrows corresponding to label "c" indicate regional transport direction in the Foothills/direction of motion of the hinterland indenter (Coastal Range plus (?) Central Range) toward the Foothills. (b) NNW-SSE seismic and geological sections (location in Figure 3a) across the Hsinchu-Hsincheng fault system. Modified after *Yang et al.* [1994]. (c) Two-stage kinematic model of interacting low-angle thrusts and reactivated normal faults in the northern part of the NW Taiwan arcuate belt. Open plane is obliquely reactivated normal fault; shaded unit is low-angle thrust sheet. Lines within the thin-skinned low-angle thrust sheet correspond to shortening trajectories.

edge of the Peikang High, as discussed in section 3.1.1 based on geological evidence (Figure 2c). The rapid decrease in earthquake occurrence at nearly 15 km depth marks the transition from an upper crustal seismogenic layer to a more ductile lower crust. In addition, major recent thrust earthquakes south of the Peikang High (Paiho, 18 January 1964; Chiali, 12 March 1991; Tapu, 15 December 1995; Chiayi, 25 September 1999; data from the Central Weather Bureau, Taiwan) are concentrated within a 13–17 km depth interval within the pre-Tertiary basement as pointed out by *Mouthereau et al.* [2001a]. This distribution argues in favor of the occurrence of an active detachment at midcrustal level, probably at the brittle-ductile transition (Figure 6a). The Chukou thrust presumably flattens at this inferred deep detachment (Figure 6a) which is required for accommodating crustal shortening and displacements related to structural inversion beneath the Coastal Plain and therefore for balancing geological cross sections through the foreland thrust belt [e.g., *Mouthereau et al.*, 2001a] (Figure 2c).

[19] In north central Taiwan, the distribution with depth of earthquakes and microearthquakes over the same period

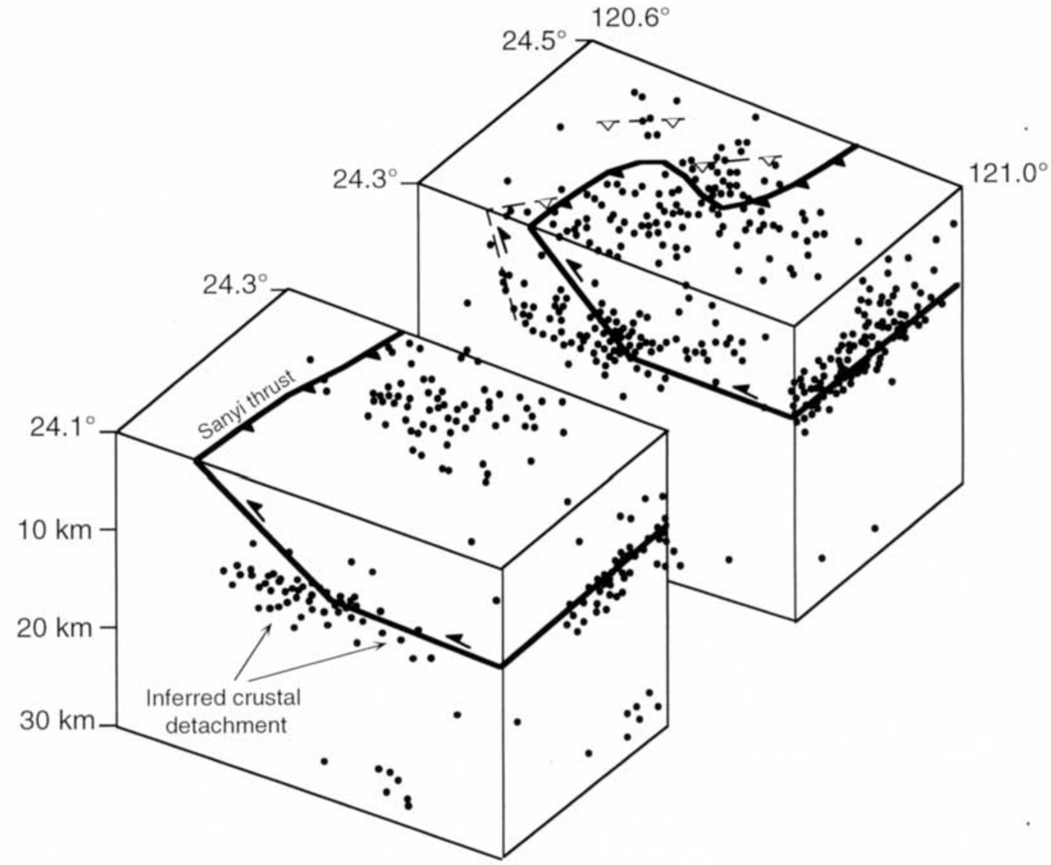


Figure 4. Distribution of microearthquakes in NW Taiwan and inferred crustal detachment at 10–12 km depth ramping up to the surface as the Sanyi thrust. Note that the inferred detachment extends beneath the Coastal Plain where it presumably accommodates the transpressional reactivation of ENE trending normal faults of the margin (dashed lines). Modified after *Lin et al.* [1989].

clearly indicates two seismogenic layers (Figure 6b), a shallower one between 2 and 14 km, and a deeper one at ~20–30 km, with an intermediate nearly aseismic zone. The upper and lower seismogenic layers probably correspond to the brittle upper crust and the brittle lowermost crust (?)-upper mantle layer, while the nearly aseismic layer may correspond to the ductile lower crust. The centroid depth of the Chi-Chi main shock determined from teleseismic body waveform is 11–12 ± 5 km [*Kao and Chen*, 2000]. The precisely relocated largest shocks of the Chi-Chi (21 September 1999) sequence ("a" mechanisms of Figure 5a) correspond to thrust events along a shallow dipping thrust zone at around 10–12 km merging as the Chelungpu thrust [*Kao and Chen*, 2000], which presumably roots (or at least flattens) at 13–17 km depth, at the top of the ductile crust which localized a deep detachment (Figure 6b) [see also *Mouthereau et al.*, 2001b]. East of the Chelungpu thrust, earthquakes show a homogeneous distribution with depth and reflect crustal collisional deformation (Figure 5b). As in Figure 5a, the 5–6 km thick cover of the western Coastal Plain shows no seismicity; however, the cluster of earthquakes west of the Chelungpu thrust trace at 8–12 km depth, which displays mixed normal and reverse type focal mechanisms ("c" mechanisms of Figure 5b), suggests that the basement deforms at depth also beneath the Coastal Plain. We propose that the normal-type focal mechanisms reflect flexural bending of the marginal crust in the footwall of the Chelungpu thrust at the time of the Chi-Chi earthquake, while reverse-type focal mechanisms support the occurrence of an incipient crustal thrust beneath the foreland (Figure 6b). As in the south central Foothills, "b" mechanisms (Figure 5a) support the right-lateral transpressional reactivation of ENE trending features both in the fold-thrust belt and beneath the foreland. As an example, the right-lateral movement along the ENE trending Tuntzuchiao fault gave rise to the 20 April 1935 (local magnitude (M_L) of 7.1) Taiwan earthquake. We propose that the crustal detachment extends beneath the fold-thrust belt and westward below the Coastal Plain and accommodates present-day structural inversion and related motion along high-angle thrusts.

[20] Summarizing, the Taiwan example illustrates that ongoing deep-seated thrusting reorganizes the geometry of

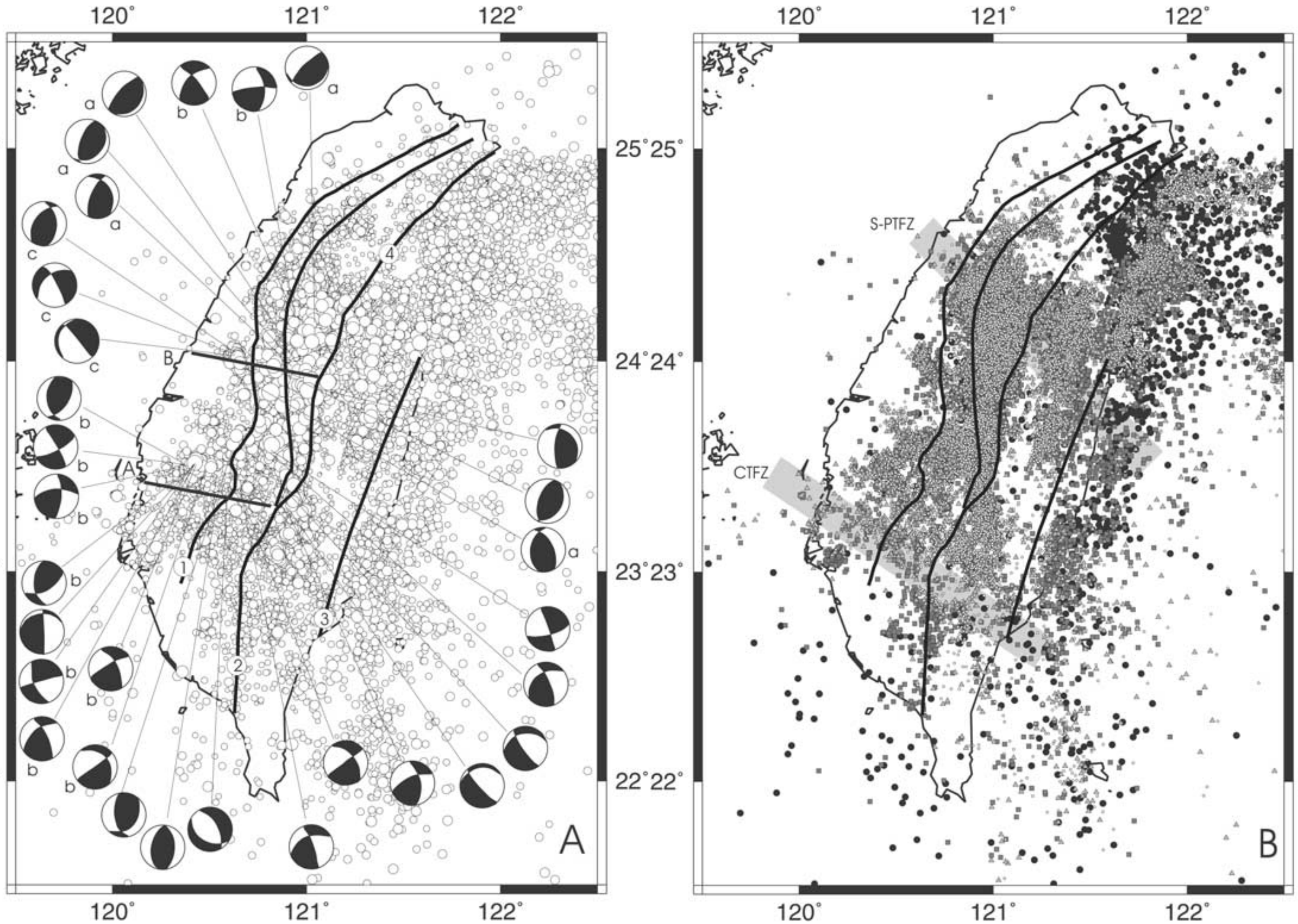


Figure 5. Seismicity in Taiwan over the period September 1999–February 2000 (data from Central Weather Bureau, Taiwan). The major structural features (1, Chukou-Chelungpu-Sanyi thrusts; 2, Chuchih-Chaochou fault; 3, Longitudinal Valley fault; 4, Lishan fault) have also been reported. (a) Distribution of earthquakes as a function of magnitude (symbols proportional to magnitude). Focal mechanisms of mainshocks and aftershocks of the Chi-Chi earthquake (21 September 1999) sequence after *Kao and Chen* [2000]. Among the total set, several mechanisms have been distinguished, which are labeled as follows: a, major subevents (thrust type) of the mainshock along the Chelungpu thrust; b, mixed reverse and strike-slip type aftershocks presumably reflecting the right-lateral transpressional reactivation of ENE trending normal faults of the margin; c, mixed reverse and normal aftershocks reflecting crustal deformation west of (and below) the Chelungpu thrust beneath the Coastal Plain. (b) Distribution of earthquakes as a function of focal depth. Small open circles correspond to 0–8 km; light shaded triangles, 8–15 km; dark shaded squares, 15–30 km; solid circles, >30 km.

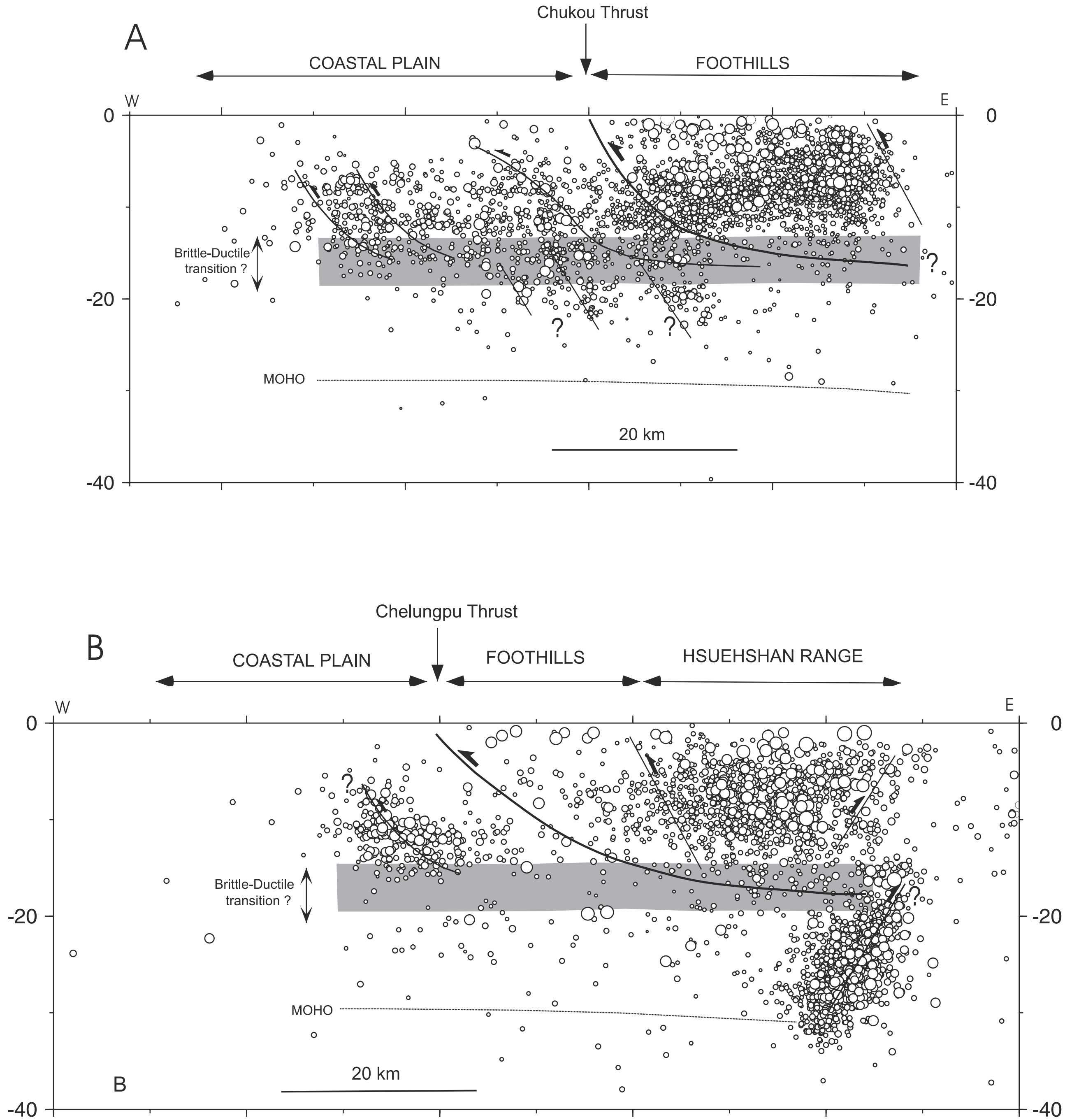


Figure 6. Cross sections through the Taiwan foreland showing depth distribution of earthquakes. Location is given on Figure 5. (a) Cross section in the Chiayi area. (b) Cross section in the Chelungpu area. Note the contrast of seismicity on both sides of the Chelungpu and Chukou thrusts that mark the western boundaries of the Foothills. The seismicity in the basement beneath the Coastal Plain presumably reflects compressional and/or flexural reactivation of preexisting normal faults inherited from the earlier extensional history of the margin (see text). The upper crust is presumably partly decoupled from the lower crust along a midcrustal weakness zone.

the orogenic wedge. This occurs through out-of-sequence basement-involved thrusts beneath the foreland [*Kao and Chen*, 2000]. The deep thrust sheets presumably detach at a depth of a ~13–17 km along crustal low-angle zones of original mechanical weakness. This depth range is consistent with the brittle-ductile transition predicted by strength envelopes calculated for a 27–30 km thick transitional crust under a 30°/km geothermal gradient [*Wu et al.*, 1997].

Structural evidence of early reactivation of normal faults of the Taihsi basin show that the crustal detachment was already active before the emplacement of the thin-skinned thrust units. As a result, the deep-seated detachment was probably active (at least) during the early and late (present-day) tectonic evolution of the Taiwan fold-thrust belt and foreland.

3.2.2. Seismotectonic Signature of Along-Strike Deep Segmentation of the Taiwan Foreland

[21] The distribution of earthquake epicenters in the Taiwan Foothills and Coastal Plain shows along-strike variations in earthquake occurrences (Figure 5b). Three main domains can be depicted west of the Lishan fault: They are bounded by the Sanyi-Puli Transfer fault zone and the Chishan Transfer fault zone [*Lacombe et al.*, 2001]. North of the Sanyi-Puli Transfer fault zone, earthquakes are few and of low magnitudes (Figure 5a); together with GPS data which show no significant westward displacement of the Taiwan Foothills with respect to South China, they indicate that this northern area presumably undergoes the latest stage of collision. In central Taiwan, between the Sanyi-Puli Transfer fault zone and the Chishan Transfer fault zone, shallow earthquakes (0–30 km) are abundant and related to collisional crustal shortening (Figure 5b). Within this central domain, however, the seismic signature of basement-involved shortening may also vary along strike, as inferred from comparison between Figures 6a and 6b. South of the Chishan Transfer fault zone, shallow earthquakes are scarce and small in magnitude (Figure 5a). A noticeable thing is the lack of shallow seismicity between 0 and 20–25 km (Figure 5b): This southern domain corresponds to the nearly aseismically deforming onland extension of the Manila accretionary wedge, which is at present-day escaping toward the SW [*Lacombe et al.*, 2001]. Seismotectonic data therefore lead to considering the Sanyi-Puli Transfer fault zone and the Chishan Transfer fault zone as major transitions which may reflect both anchoring of the Taiwan collision and segmentation of the belt at the crustal (lithospheric?) scale.

4. Basement Thrusting and Deep Detachment Tectonics in the Neogene Foreland Thrust Belt of the Western Alps

4.1. Structural Evidence

[22] The foreland thrust belt of the Western Alps extends from Switzerland to the Mediterranean. It mainly comprises the Jura Mountains, the Helvetic domains, and the Subalpine chains (Figure 7a). These fold-thrust belt segments developed mainly during the late Miocene in response to crustal shortening because of the underthrusting of the European lithosphere.

[23] The Subalpine fold-thrust belts are separated from the stable foreland by the Molasse Basin which extends southward to the eastern border of Leman. The Jura fold-thrust belt developed on the external side of the Molasse foredeep as the outermost and most recent fold belt segment (Figure 7a). Stratigraphic and paleontological data indicate that folding and thrusting of the Jura was a short-lived event, probably between 12 and 3.3 Myr ago [*Burkhard and Sommaruga*, 1998; *Becker*, 1999]. The structural grain of the curved belt swings a full 90° from a N-S direction at the SW to an E-W direction at the NE, and designs a clear salient (Figure 7a).

[24] To the first order, the thrust front of the Subalpine chains follows the regional bend of the western Alpine arc; in detail, the shape of this thrust front reflects differential advancing of the thin-skinned foreland thrust belt segments onto the stable foreland, in response to lateral variations of the pre-orogenic Mesozoic cover thickness and of the distribution of the potential detachment levels, such as the Triassic evaporites which constitute the main basal low-friction cover horizon in the Jura [e.g., *Philippe et al.* 1996; *Sommaruga*, 1997, 1999] or also the Middle Jurassic or Lower Cretaceous shales.

[25] The basement is also involved in shortening in the Alpine foreland (Figure 7a). The External Crystalline Massifs (Aiguilles Rouges, Mont Blanc, Aar, Belledonne-Pelvoux) underwent an early uplift history during the early Oligocene phase of the Mesoalpine events: For instance, the Mont Blanc-Aar nappes started to take shape while the formation of the major Helvetic nappes began [*Escher et al.*, 1997]. Tectonic inversion of previous normal faults likely played an important role during the early-formed Helvetic units. The Aar Massif was then uplifted during Oligocene–early Miocene times, and together with the Aiguilles Rouges/Mont Blanc Massifs, it finally underwent additional shortening and uplift during middle and late Miocene, coeval with shortening in the Subalpine Molasse and the Jura Mountains [*Pfiffner et al.*, 1997a]. The thrust fault in the northern flank of the Aar Massif, which puts the basement onto the autochthonous Mesozoic foreland cover, is likely to have formed late in the local sequence in relation to shortening within the Subalpine Molasse, and could be partly responsible for the late uplift of the Massif. This late bulge of the Aar Massif, which caused refolding of the basal thrust of the previously emplaced overlying Helvetic nappes, is kinematically linked to the deformation in the foreland through a basement detachment. It is imaged by seismics, at a depth of nearly 5 km beneath the transition Molasse Basin–Folded Jura and is referred to as the Jura Detachment [*Pfiffner et al.*, 1997b]. This detachment deepens eastward to ~10 km beneath the transition between the Prealps and the Molasse Basin, then to a depth of 20 km below the Aar-Gotthard Massifs and presumably roots at the upper-lower crust transition [*Pfiffner and Heitzmann*, 1997; *Pfiffner et al.*, 1997b]. Such a detachment in the basement is required to accommodate displacements and shortening related to inversion of Permo-Carboniferous basins recognized below the Molasse Basin [e.g., *Philippe*, 1995; *Pfiffner et al.*, 1997b]. As a result, the most probable mechanism for the tectonic evolution of the Molasse Basin–Jura Mountains system comprises both thin-skinned tectonics along a basal décollement within the Triassic evaporites [e.g., *Philippe et al.*, 1996] and basement involvement along a deeper detachment which

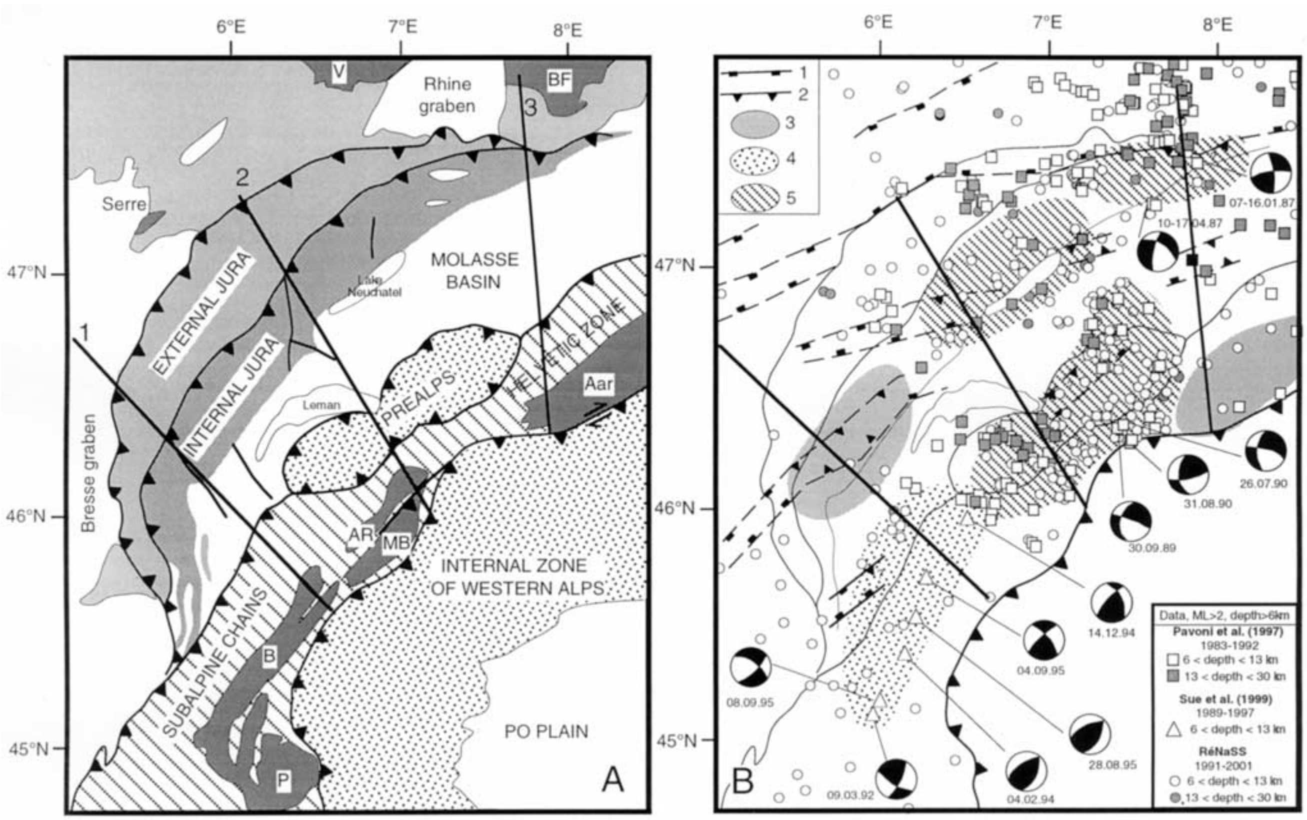
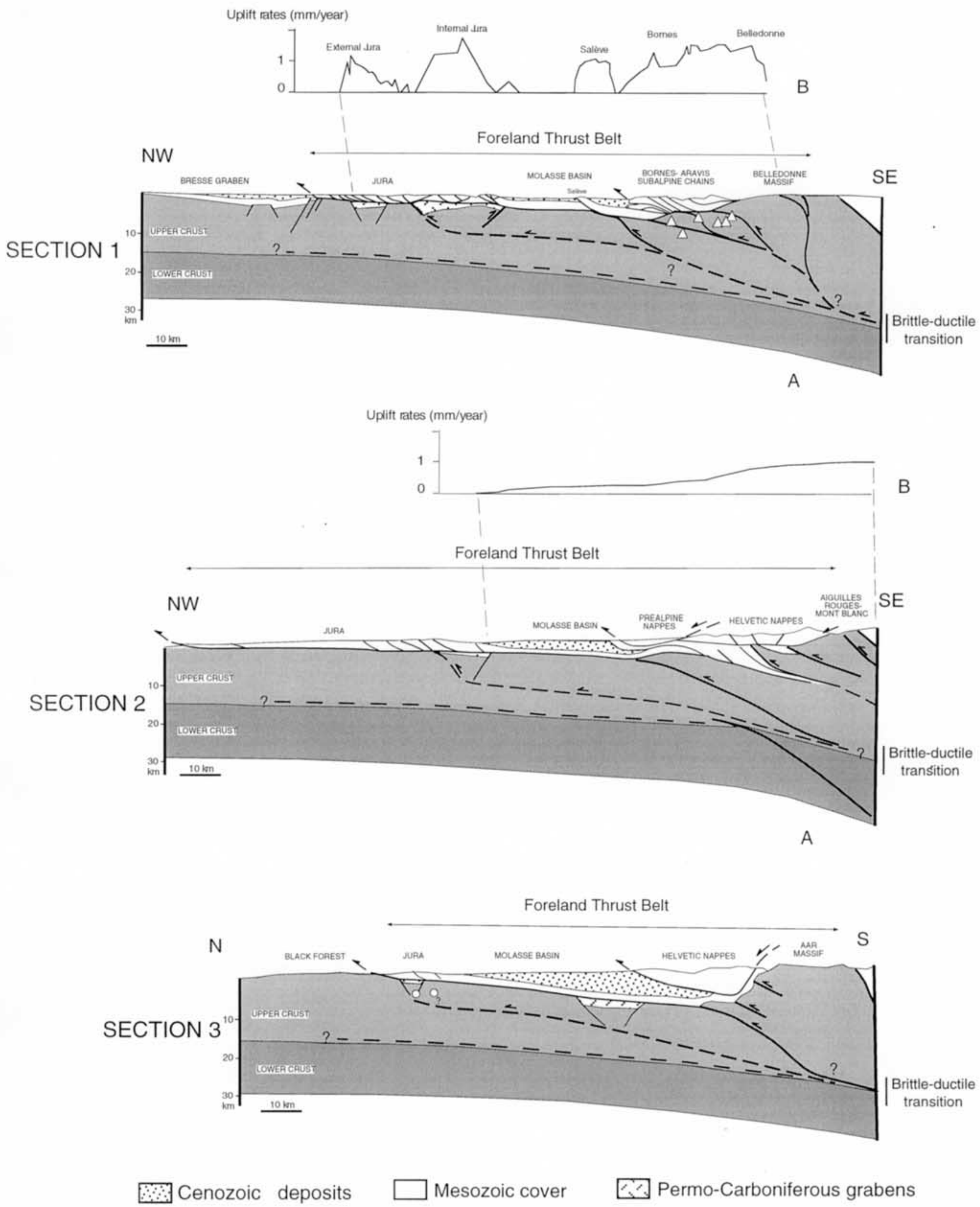


Figure 7. Simplified structural map of the NW Alpine foreland. (a) Lines numbered 1, 2, and 3 correspond to cross sections of Figure 8. Crystalline massifs are abbreviated as follows: V, Vosges; BF, Black Forest; AR, Aiguilles Rouges; MB, Mont Blanc; B, Belledonne; P, Pelvoux. Seismicity of $M_L > 2$ and focal depth >6 km in the NW Alpine foreland. (b) Numbered patterns are as follows: 1, normal fault bounding underlying Permo-Carboniferous basin; 2, reactivated normal fault/inverted Permo-Carboniferous basins; 3, areas of present-day basement-involved shortening inferred from high present-day uplift rates [e.g., *Jouanne et al.*, 1995; *Gubler et al.*, 1992; *Kahle*, 1997]; 4, areas of present-day basement-involved shortening inferred from both high present-day uplift rates and seismicity; 5, areas of present-day basement-involved shortening inferred from seismicity. Note that Neogene-present-day basement-involved shortening occurs mainly in areas where Permo-Carboniferous basins underlie the detached Mesozoic cover.

accounts for inversion of underlying Permo-Carboniferous and accommodates stacking of basement thrust units beneath the External Crystalline Massifs.

[26] Deep seismic lines across the Belledonne Massif and Subalpine chains also suggest that the basal Belledonne thrust ramps up from a depth of ~12 km under the Belledonne and Bornes Chains to 7 km under the Bornes-Molasse Basin boundary [e.g., *Mugnier et al.*, 1990]. The Aiguilles Rouges-Mont Blanc Massifs show basement imbricates with fault displacements possibly reaching 8–10 km along the basal thrust. The geometry at depth of the basement thrusts imaged by seismics suggests a detachment at 10–12 km depth. Also, thrust faulting in the Aiguilles Rouges Massif might have been facilitated by preexisting Permo-Carboniferous graben structures [*Pfiffner et al.*, 1997a]. We therefore tentatively propose that a deepening eastward detachment occurs within the upper crust beneath the Molasse Basin, the Subalpine chains, and the External Crystalline Massifs and that it roots at the brittle-ductile transition (Figure 8, sections 1, 2, and 3).

[27] Although early inversion of underlying Permo-Carboniferous grabens cannot be definitely ruled out, structural and seismic data rather suggest that involvement of the crystalline basement began recently in the outermost parts of the Alpine foreland thrust belt, either during the latest stage of frontal thin-skinned thrusting or just after it ceased. Local basement elevation within the southern Jura Mountains [*Laubscher*, 1986; *Guellec et al.*, 1990] presumably reflects recent basement imbricates [*Roure et al.*, 1990; *Philippe*, 1995] rather than a basement horst inherited from the Oligocene extension. Basement uplift possibly occurred in response to the (oblique) inversion of an underlying Permo-Carboniferous basin (Figure 7b): This uplift postdated the westward translation of the detached Mesozoic series and therefore caused deformation of the allochthon [*Roure and Colletta*, 1996]. We


Uplift rates (mm/year)
External Jura
Internal Jura
Salève
Bornes
Belledonne
B
Foreland Thrust Belt
NW
SE
BRESSE GRABEN
JURA
MOLASSE BASIN
BORNES- ARAVIS SUBALPINE CHAINS
BELLEDONNE MASSIF
UPPER CRUST
LOWER CRUST
SECTION 1
10 km
Brittle-ductile transition
A
Uplift rates (mm/year)
B
Foreland Thrust Belt
NW
SE
JURA
MOLASSE BASIN
PREALPINE NAPPES
HELVETIC NAPPES
AIGUILLES ROUGES-MONT BLANC
UPPER CRUST
LOWER CRUST
SECTION 2
10 km
Brittle-ductile transition
A
Foreland Thrust Belt
N
S
BLACK FOREST
JURA
MOLASSE BASIN
HELVETIC NAPPES
AAR MASSIF
UPPER CRUST
LOWER CRUST
SECTION 3
10 km
Brittle-ductile transition
Cenozoic deposits
Mesozoic cover
Permo-Carboniferous grabens

propose that this outermost basement uplift be related to a basement high-angle thrust initiated from a reversed normal fault and which roots at the previously identified crustal detachment. Ongoing basement uplift (and therefore present-day activity of the basement flat-ramp thrust system) is further demonstrated by precise leveling in the French Alps and the Jura [*Jouanne et al.*, 1995] which indicates regional positive vertical displacements in the Subalpine chain (Belledonne, Aravis, external Bornes), in the Salève area (in the Molasse basin), and in the internal Jura (Figure 8, section 1A).

4.2. Seismotectonic Evidence

[28] As in Taiwan, we have combined recent seismotectonic data with structural data at the regional scale (Figure 7a) in an attempt at drawing a general picture of the recent deep deformation of the NW Alpine foreland. Earthquake distribution with depth helps correlate areas where Neogene basement-involved shortening has been documented on the basis of structural evidence.

[29] Seismotectonic analyses as well as in situ stress measurements support that a structural deformation mode with basement-involved shortening prevails today in the southeastern Jura belt. Contemporary near-surface stress data suggest that active thin-skinned tectonics for the Jura Mountains has ceased [*Becker*, 1999]: The recent stress field does not agree with paleostress orientations related to Jura emplacement as deduced from fault slip data [*Tschanz*, 1990; *Homberg et al.*, 1999]. Stress provinces defined in the Jura Mountains on the basis of homogeneous present-day maximum horizontal stress orientations extend far into the nondetached foreland, without signs of variations across the border between the folded and thrust Jura and its undeformed foreland. These results together indicate that the present-day tectonics has characteristics which differ from those prevailing during the main stage of Jura folding and thrusting.

[30] Seismic activity testifies that the whole crust is actively deforming in the entire internal Jura-Molasse basin domain, as proposed by *Pavoni et al.* [1997], and is not at all limited to the Mesozoic cover. This deep basement tectonics is in good agreement with occurrence of the recent and still active basement thrusting proposed by *Mosar* [1999] beneath the klippen of the Swiss Prealps in the Swiss Alps (Figure 8, section 2). The western Alpine foreland thrust belt shows some seismicity [*Deichmann*, 1992; *Deichmann and Baer*, 1990; *Pavoni et al.*, 1997; *Sue et al.*, 1999; this paper] which decreases westward and southwestward (Figure 7b). Earthquake hypocenters are distributed throughout the entire depth range of the crust down to 30 km, with a higher density in the upper crust above the brittle-ductile transition at around 15–20 km. The presence of earthquakes in the Swiss lower crust has been related to high fluid pressures [*Deichmann*, 1992]. Numerous earthquakes occur at depth shallower than 5 km, as, for instance, in the Geneva region where focal depths of events between 3 and 5 km place them within the cover (e.g., the Epagny earthquake, 15 July 1996; $M_L = 5.3$, marking the activity along the major Vuache left-lateral strike-slip fault, occurred at a depth of 3 km [*Thouvenot*, 1998; see also *Sambeth and Pavoni*, 1988]) and maybe at the cover-basement transition. In order to highlight basement tectonics in an unambiguous way, we did not consider the entire seismicity in our study. Rather, we focus on events with local magnitude larger than 2 which can reliably be considered as occurring in the basement, i.e., with focal depths >6 km. An overall picture of the present-day deformation of the basement in the area studied is given by the map of the epicenters of the earthquakes recorded by the French ReNaSS (Réseau National de Surveillance Sismique) over the period January 1991–March 2001 (Figure 7b). However, these earthquakes have not been precisely relocated, so that uncertainties may occur especially on focal depths. We consequently reported on the same map more accurately depth-located earthquake data from *Pavoni et al.* [1997] over the period 1983–1992 and from *Sue et al.* [1999] over the period 1989–1997 selected on the same criteria ($M_L > 2$; focal depth >6 km), and used them to draw structural conclusions.

[31] The map of Figure 7b shows two types of earthquakes: those north of 47°N30 and between 7°E and 8°E are more or less aligned along the border faults of the Rhine graben and clearly reflect their present-day tectonic activity and kinematics in response to Alpine shortening. South of 47°N30, the earthquakes rather mark the present-day tectonics of the outer Alpine domain. Several focal mechanisms of earthquakes from *Pavoni et al.* [1997] and *Sue et al.* [1999] are illustrated (Figure 7b): These fault plane solutions evolve from reverse-strike-slip types in the SW part (Subalpine chains) of the Alpine foreland thrust belt to strike-slip-normal types in the NE part (Helvetic nappes and eastern Jura). The related stress orientations in the Molasse Basin and Internal Jura show a fan-shaped distribution of maximum compressional horizontal stress evolving from a N-S orientation to the east associated with subperpendicular horizontal extension progressively to a NW-SE orientation to the west in the Salève-southern Jura area. In the French Subalpine chains, the derived maximum compressional horizontal stress is rather directed WNW-ESE. Earthquakes above (not reported here) and below the detachment horizon show similar focal mechanisms, which are in good agreement with present-day in situ stresses in the southern Jura

Figure 8. (opposite) Crustal-scale cross sections through the NW Alpine foreland (location on Figure 7) emphasizing basement thrusting and occurrence of a deep crustal detachment rooting at the brittle-ductile transition. Section 1-A is cross section from the Bresse graben to the Belledonne massif; section 1-B is present-day uplift rates deduced from leveling (data after *Jouanne et al.* [1995]). Open triangles correspond to earthquake data by *Sue et al.* [1999] (Figure 7b). Section 2-A is cross section from the Western Jura front to the Aiguilles Rouges/Mont Blanc massifs (modified and completed after *Mosar* [1999]). Section 2-B is present-day uplift rates deduced from leveling [data after *Gubler et al.*, 1992; *Kahle*, 1997]. Section 3 is cross section from the Black Forest to the Aar massif. White circles correspond to the 07–16 January 1987 and 10–17 April 1987 earthquake clusters reported in Figure7b [*Deichmann and Garcia-Fernandez*, 1992].

area [*Becker*, 1999], indicating similar stress orientations in the cover rocks as well as in the basement. In addition, the highest magnitudes of near-surface present-day stresses in the whole Jura Mountains are determined only a few kilometers south of the epicenter of the Clairvaux earthquake interpreted as a thrust event [*Pavoni*, 1977]. The similarity between deep and near-surface stresses, the rapid uplift of the southern internal Jura, and crustal seismicity support present-day active tectonics involving the basement below the main basal ductile horizon in the Molasse Basin and the southern Jura mountains.

[32] A striking point outlined by Figure 7b is the close relationships between the occurrence of basement thrusting and the presence of inverted Permo-Carboniferous basins beneath the Mesozoic cover. Permo-Carboniferous grabens are found in the folded Alpine cover-basement nappes, but have also been recognized by seismics and drilling in the outermost Alpine domain [*Truffert et al.*, 1990; *Gorin et al.*, 1993; *Philippe*, 1995; *Signer and Gorin*, 1995; *Diebold and Noack*, 1997; *Pfiffner et al.*, 1997b] (Figure 7b), some of them having been inverted during the Alpine shortening. The seismicity map of Figure 7b show that earthquakes often (although not always) occur in areas where Permo-Carboniferous basins have been identified beneath the detached cover (Figures 7b and 8); even in areas where seismic data do not support the actual inversion of these Permo-Carboniferous basins, clusters of earthquakes can be considered as markers of their ongoing (incipient) inversion, since it is often easier to reactivate preexisting crustal weaknesses than to initiate new faults. As a consequence, the cluster of earthquakes which occurred during the periods 07–16 January 1987 and 10–17 April 1987 at 6–7 km depth beneath the front of the Jura Mountains in northern Switzerland [*Deichmann and Garcia-Fernandez*, 1992], presumably reflects triggering and localization of basement shortening by the underlying Permo-Carboniferous graben (Figure 8, section 3), and maybe even its ongoing (forthcoming) inversion. In a similar way, and to keep internal consistency to structural interpretations, we tentatively propose in section 2 of Figure 8 the ongoing inversion of the Permo-Carboniferous graben below the transition between the Molasse Basin and the Internal Jura, as already suggested by *Mosar* [1999] (Figure 7b).

[33] To summarize, deep-seated thrusting occurred (during and) after the late evolution of the outermost part of the Jura foreland thrust belt, after cover detached along the basal Triassic thrust detachment. Deep-seated thrusting caused basement uplift and late deformation of the thin-skinned nappes. A more evolved stage of refolding of previous shallow nappes in response to deep thrusting is seen in the Crystalline Massifs and Helvetic domains where basement culminates. The tectonic basement window of Belledone results from basement imbricates that induced refolding of the older shallow thin-skinned thrust sheets of the Subalpine Chains and the Prealpine nappes. In a nearly similar way, the Mont Blanc-Aiguilles Rouges Massifs uplift deformed the Morcles nappe fold structures [e.g., *Pfiffner*, 1993]. Number of basement thrusts presumably nucleated from preexisting normal faults bounding Permo-Carboniferous basins. The late Miocene and still active uplift of the External Crystalline Massifs represent the last stage of a longer uplift history of the shortened basement which started as soon as (Eocene)-Oligocene times, before and partly coeval with emplacement of nappes detached along ductile basal horizons. Basement stacking occurred along thrusts rooting into an upper crustal detachment which deepens orogenward down into the brittle-ductile transition. Decoupling at the brittle-ductile transition probably became efficient, while the underthrusted European crust underwent progressive ductilization, which allowed generation of successive basement thrust units and nappes. We propose that the detachment extends beneath the outermost foreland and that its (local?) activation helped accommodating shortening and basin inversion in the Alpine far foreland.

5. Basement Thrusting and Deep Detachment Tectonics in the Paleogene North Pyrenean-Provençal Foreland

[34] The Pyrenees belt formed in Late Cretaceous–Eocene times in response to the collision between the European plate and the Iberian-Sardinian-Corsican block [*Arthaud and Séguret*, 1981]. The belt shows a prominent bend in southeastern France, then it extends farther east in Provence and more or less connects to the western Alps (Figure 9). Deep seismic profiling [*Etude Continentale et Océanique par Reflexion et Refraction Sismique Pyrenees Team*, 1988] shows that the belt displays a deep asymmetric structure resulting from the underthrusting of the flexed Iberian continental lithosphere below the Eurasian lithospheric mantle indenter. A major decoupling occurs at the crust-mantle boundary of the subducting lithosphere. Shortening was accommodated within the underthrust Iberian plate by deep crustal thickening and by major back thrusting on top of the indenter, thus designing a lithospheric-scale triangle structure. The resulting surficial structure consists of an asymmetric double verging crustal wedge, with the elevated axial part of the Pyrenees belt evolving into opposite verging conjugate foreland fold-thrust belts made of allochthonous cover and basement units; at their front, the Ebro and Aquitaine foreland basins developed.

[35] In contrast, the structure of the eastern Provençal segment is less well constrained, mainly because the later Oligo-Miocene extension and opening of the Liguro-Provençal basin concealed part of the Pyrenean belt, thus truncating the structural continuity between Pyrenees and Provence and flooding the inner part of the Provençal belt under the Gulf of Lions. In addition, the foreland of the Provence segment is poorly defined and generally diffuse. The complex array of multidirectional compressional structures is related to the widespread structural inversion of Paleozoic and Mesozoic basins during the Eocene "Pyrenean" and the Miocene "Alpine" shortenings [*Roure and Colletta*, 1996], which prevented typical foreland flexural evolution. The absence of thick synorogenic to postorogenic series makes the Provençal segment a suitable (although

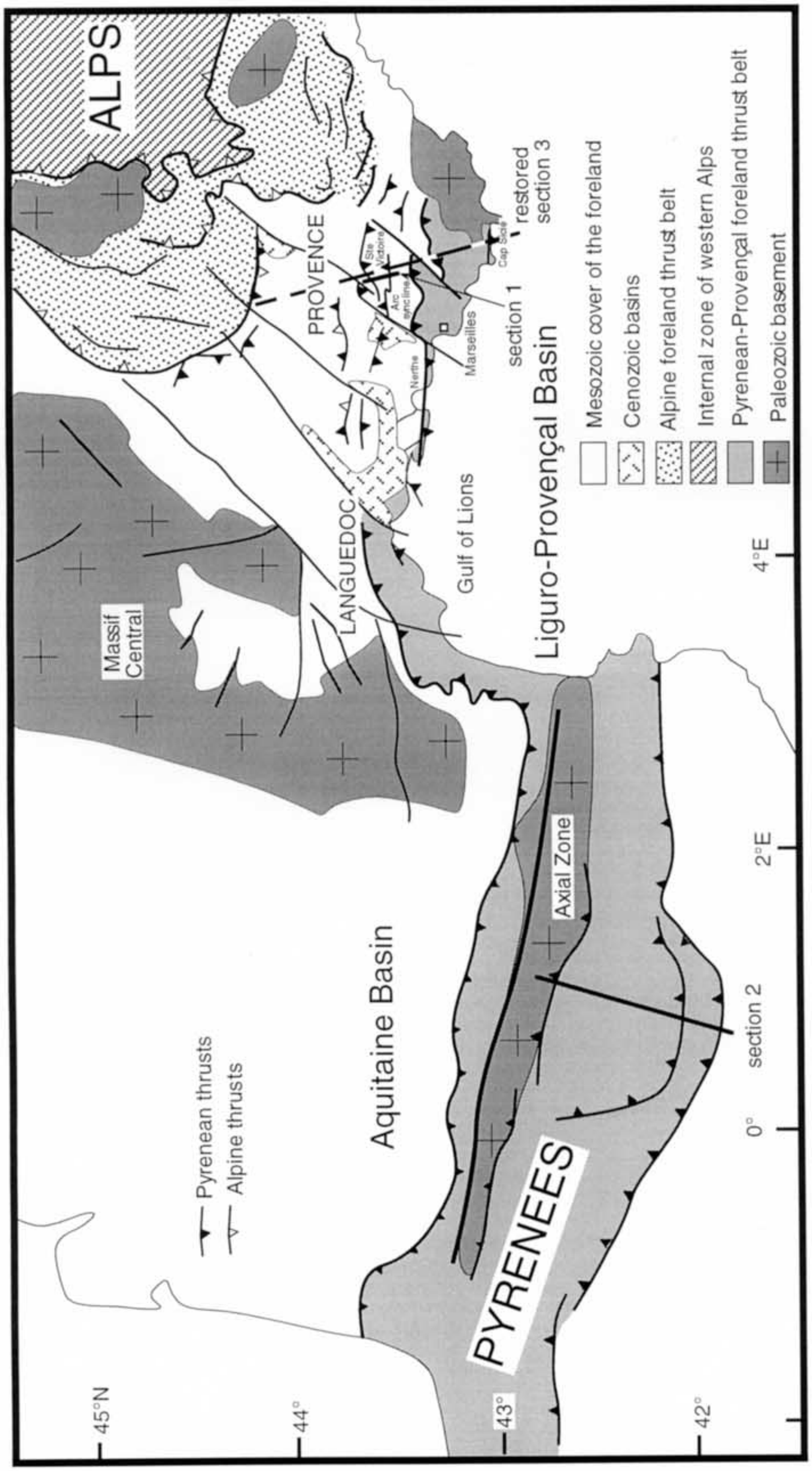


Figure 9. Simplified structural map of the Pyrenean-Provençal belt in southern France. The thick lines correspond to the location of the present-day sections of Figures 10a and 10b; the dashed line shows the location of the palinspastic section shown in Figure 10c.

complex) area to investigate basement fault reactivation at the front of a young but no longer active orogen.

[36] In Provence, the Pyrenean shortening occurred from the Late Cretaceous to late Eocene [e.g., *Tempier and Durand*, 1981; *Lacombe et al.*, 1992]. E-W trending folds and thrusts developed within the Mesozoic cover detached from the underlying basement above the Triassic evaporites [e.g., *Tempier*, 1987]. The front of the allochthonous Provençal nappes connects westwards to the front of the Languedoc nappes and farther west to the northern Pyrenean front (Figure 9). In the southernmost onshore areas, the Paleozoic basement is clearly involved in shortening, as in the Cap Sicié and the Nerthe structures (Figure 9) where crystalline rocks were thrust onto Permian and Mesozoic formations during the Eocene [*Arthaud and Séguret*, 1981]. A similar basement involvement has been documented in neighboring segments of the Pyrenean belt; for instance, the Nogueras basement window in the Pyrenees results from basement imbricates that refold the Tremp units [*Roure et al.*, 1990] (Figure 10b). In Languedoc, the basement high at the rear of the Corbieres nappe which now lies offshore within the Gulf of Lions margin was presumably created in the same way [*Vially and Tremolieres*, 1996], suggesting similar basement-cover relationships than in the Western Alps.

[37] At the front of the Provençal allochthonous nappes, number of geological evidence support widespread Cenozoic structural inversion, but it is sometimes difficult to unambiguously discriminate between inversion of preorogenic Mesozoic extensional basins and reactivation of deep basement faults, and between the Eocene Pyrenean and the Miocene Alpine shortenings which can be coaxial in some places. We retain hereinafter the example of the Sainte-Victoire Range for which reactivation of a basement fault in response to Pyrenean stresses can be documented with some confidence.

[38] Inversion of preexisting basement normal faults accounts well for the abnormally high elevation of isolated structures in the foreland, far from the orogenic front. An example of this is provided by the Sainte-Victoire range in Provence. It consists of an E-W trending, 1000 m high fold-thrust range, made of folded Jurassic limestones and which displays along-strike variations of the main thrust vergence, from forward (northward) directed thrusting to the east to back thrusting (southward) to the west, with a double verging pop-up structure in its central part (Figure 9). The south verging thrust unit overlies thick synkinematic breccias and Late Cretaceous to Eocene continental sediments of the Arc syncline (Figure 10a).

[39] Previous interpretations have considered the Sainte-Victoire Range as a thin-skinned feature forming part of the north verging Pyrenean-Provençal nappes [e.g., *Tempier*, 1987]. Geophysical data indicate a southward deepening of the basement below the range, and thick Permian strata have been identified along the eastern margin of the Arc syncline, whereas the Triassic evaporites which constitute the regional thrust detachment level rest directly on the basement north of the Sainte-Victoire. This suggests that the range is superimposed on a south dipping Permian high-angle fault limiting an underlying late Paleozoic basin (Figure 10a). On the basis of sandbox simulations of basement-controlled inversion, *Roure and Colletta* [1996] proposed that the Sainte-Victoire structure can be balanced in situ, most of the shallow deformation being balanced by an equal amount of basement shortening. The alternative hypothesis is that the entire Arc syncline structure is detached from the substratum above the Triassic thrust detachment and that the inferred basement fault has simply localized deformation within the detached cover. As suggested by the cross section of Figure 10a, the inversion of the basement structure (or even its passive existence in the second scenario) probably caused stress concentrations in the overlying cover and induced activation of shallow intracover potential décollement levels, such as the regional Triassic evaporites, but also the upper Jurassic shales.

[40] Occurrence of syntectonic proximal breccias found at the southern front of (or locally thrust by) the Sainte-Victoire unit provides the opportunity to accurately establish the timing of deformation above an inverted basement fault through sedimentary dating of fold development. Two generations of breccias can be recognized [*Tempier and Durand*, 1981; *Lacombe et al.*, 1992] (Figure 10a): the first one, Campanian in age, unconformably overlies the vertical Jurassic strata of the Bimont fold, suggesting an early stage of folding before the development of the main fold-thrust structure. The second one, Dano-Montian in age, unconformably overlies both the previous breccia and the Jurassic strata of the Bimont fold. Both of them have been thrust by the main Jurassic unit, during the main Eocene tangential phase and the coeval inversion of the underlying Permian basin. Additional paleostress reconstructions based on tectonic analysis of minor fault sets [*Lacombe et al.*, 1992] show that between the Late Cretaceous and the late Eocene, strike-slip fault regimes have regionally prevailed during the Pyrenean compression (Figure 10a), except during the early stages of fold development associated with deposition of the Campanian and Dano-Montian synkinematic breccia, and during the late Eocene tangential event. This strongly suggests that away from the thrust belt, strike-slip stress regimes prevailed, but that stress permutations (principal vertical stress $\sigma 2$ switching to vertical $\sigma 3$) occurred locally during early folding and main thrusting in response to stress concentrations by the likely reactivated deep fault.

[41] The widespread occurrence of strike-slip stress regime within the Pyrenean foreland during the Late Cretaceous to late Eocene Pyrenean compression has been discussed by *Lacombe* [2000], on the basis of other regional works [*Letouzey*, 1986; *Bergerat*, 1987; *Bles et al.*, 1989; *Lacombe et al.*, 1992, 1996]. From a regional point of view, the Pyrenean compression of the Alpine phase resulted in a continuous shortening from the Late Cretaceous to the early Oligocene marked in the foreland by strike-slip faults, in agreement with numerical models [*Sassi et al.*, 1993; *Sassi and Faure*, 1997]. More or less synchronous tectonic pulses (Campanian-Maestrichtian, Paleocene, late Eocene) marked by isolated folding or reverse faulting in the foreland and nappe emplacement in the north and south Pyrenean fold-thrust belts have, however, been identified through careful tectonosedimentary analyses [e.g., *Déramond et al.*, 1993;

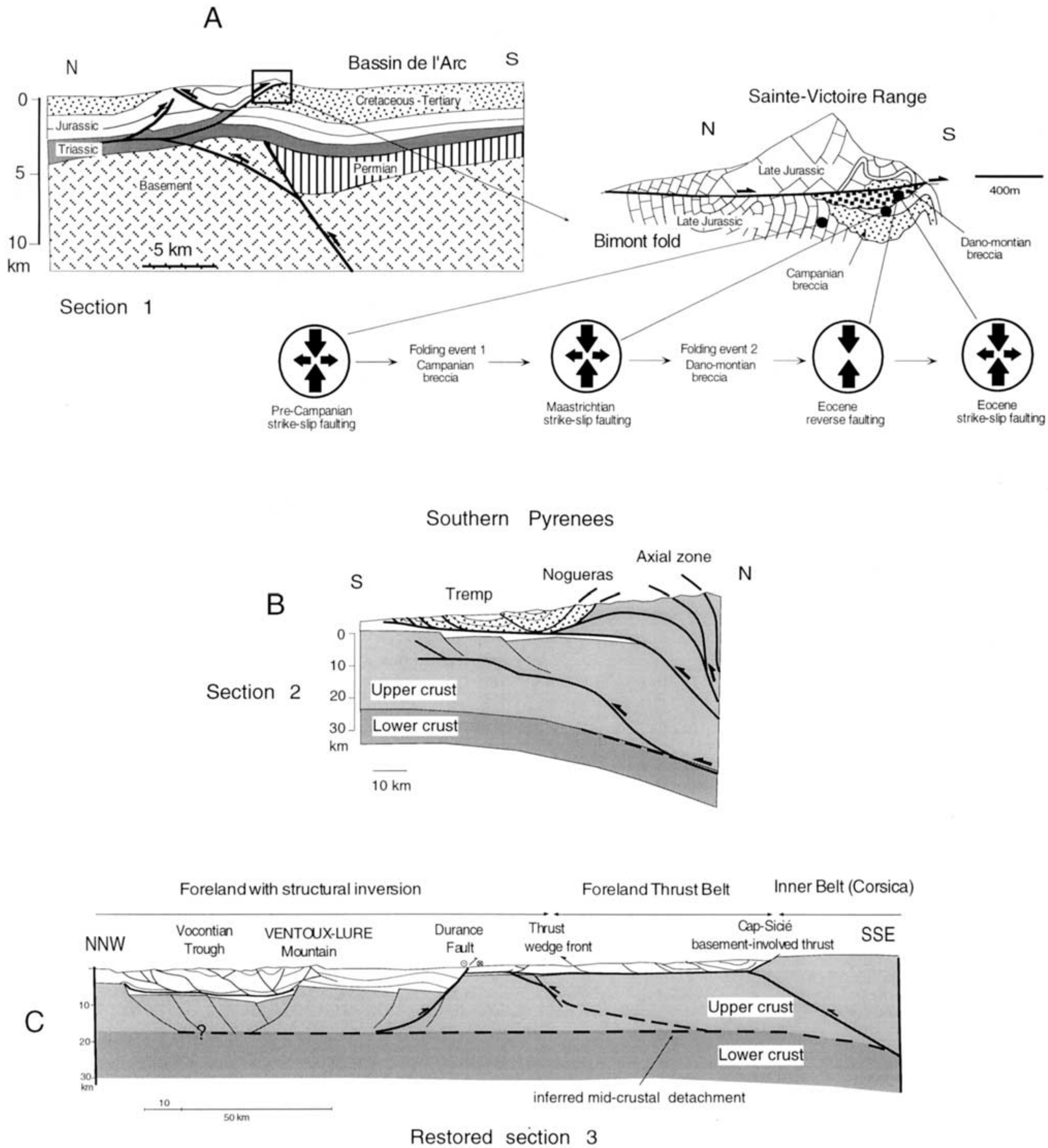


Figure 10. (a) Present geological cross section through the Sainte-Victoire Range emphasizing the inversion of an underlying Permo-Carboniferous basin (modified after *Roure and Colletta* [1996]) and detailed cross section of the Sainte-Victoire Range showing the synkinematic Dano-Montian and Campanian breccia thrust by the Sainte-Victoire Mesozoic unit and time distribution of successive stress regimes (large arrows show σ1 axes; small arrows show σ3 axes in map view) and folding events during the Pyrenean shortening. (b) Present crustal-scale geological cross section through the Nogueras area (location on Figure 9) showing refolding of shallow thin-skinned nappes and basement uplift in response to deep-seated thrusting (modified after *Roure et al.* [1990]). (c) Restored crustal-scale cross section through the Provençal foreland thrust belt at the end of the Pyrenean orogeny emphasizing basement thrusting and occurrence of superimposed shallow and deep detachment tectonics (modified after *Vially and Tremolieres* [1996]).

Rocher et al., 2000]. These tectonic pulses presumably reflect periods of increasing coupling between the orogen and the foreland during the Iberia-Eurasia collision within the framework of the Africa-Eurasia convergence. During these periods of increasing coupling at the crustal scale, potential deep detachments were probably activated in addition to shallow cover detachments, and allowed inversion of basement structures far in the foreland.

[42] The section of Figure 10c cuts through the assumed geometry of the Provençal domain at the end of the Pyrenean orogeny and before the Oligo-Miocene extension associated with the opening of the Provençal basin. To this purpose, the extensional reactivation of earlier Pyrenean compressional features as normal faults or low-angle detachments has been tentatively restored. The section outlines the reactivation of basement faults in response to Pyrenean stresses up to 50 km away from the present-day thrust wedge front, and maybe farther north (inversion of the Vocontian Trough [*Vially and Tremolieres*, 1996]). The section further assumes that the shortened upper crust was decoupled from the lower crust by a detachment at mid-crustal level. *Lacombe and Mouthereau* [1999] argued that such a potential deep detachment may extend northward because it is required to accommodate displacements, although moderate, associated with within-plate Pyrenean compressional deformation in the far foreland, as in the French Massif Central [*Bles et al.*, 1989] and the Paris Basin [*Lamarche et al.*, 1998; *Lacombe and Obert*, 2000].

6. Discussion and Conclusion: Occurrence and Timing of Basement Fault Reactivation and Deep Detachment Tectonics in Forelands of Orogens

[43] As discussed in section 5, the North Pyrenean foreland provides evidence for changes through time of the mechanical coupling between the foreland and the orogen. More generally, the timing and intensity of foreland compressional deformation indicate that mechanical coupling between an evolving orogen and its foreland can vary significantly. Processes controlling such a coupling are poorly known, but it can reasonably be assumed that apart from relative plate convergence rates and directions, important parameters are the crustal strength and configuration of the collided continental margin, including thickness and buoyancy, rheological and thermal structure, sediment supply, geometry and pattern of discontinuities (see detailed discussion by *Ziegler et al.* [1998]). Occurrence of deformations in the foreland far from the orogen implies that the plate was sufficiently rigid to transmit orogenic stresses over considerable distance in the foreland, but at the same time, it requires that this plate as a whole was not completely rigid and allowed progressive activation of deep crustal detachments.

6.1. Basement Fault Reactivation in Forelands of Orogens and Far-Field Orogenic Stresses

[44] Within-plate compressional deformations basically require that the buildup of intraplate compressional stresses, which in most cases mainly result from a (far-field) transmission from the collisional plate boundary, leads to sufficient stress magnitudes to overcome the local strength of the crust and cause reactivation of preexisting weaknesses. Reactivation of basement faults in forelands therefore requires at the same time available well-oriented discontinuities and a good transmission (and/or a low attenuation) of orogenic stresses, the latter depending on the amount of coupling between the orogen and the foreland. In the north Pyrenean foreland, the magnitude of the deviatoric component of the N-S trending maximum horizontal principal stress (σ1D) has been evaluated in the Mesozoic cover based on calcite twin analyses [*Lacombe et al.*, 1996]. An exponential decrease of this value with increasing distance to the orogenic front has been evidenced, as observed for differential stresses at the front of the Appalachian and Sevier orogens [*Van der Pluijm et al.*, 1997]. At 100 km from the front, the average σ1D is $\sim$50 MPa; it decreases to 30–40 MPa at 500–700 km away from the front.

[45] These stress values were sufficient to induce reactivation of preexisting (basement) faults within the foreland since frictional resistance is generally less than shear rupture strength under the same confining pressure [e.g., *Etheridge*, 1986], so that the stress necessary to initiate sliding on favorably oriented preexisting faults is less than that needed to initiate new faults in intact rocks. In addition, depending on the nature of fractured rocks, preexisting fractures may become overpressured during compression, thereby decreasing the effective normal stress holding the opposite walls of the faults together [*Sibson*, 1993]. Preexisting faults therefore act as crustal weaknesses when the crust undergoes later shortening [*Cooper and Williams*, 1989]. At a more regional scale, the high fluid pressure which may prevail in sedimentary basins depending on specific basinal conditions such as high sedimentation rates and the presence of sealing beds can be such that the local strength is considerably lowered. Even very small variations in magnitudes of applied far-field stresses are sufficient to induce reactivation. This point has two consequences of considerable importance: First, stresses cannot be transmitted through sedimentary high fluid pressured domains, and therefore they concentrate in high-strength areas of the crust where they can reach crustal strength [*Zoback et al.*, 1993]; second, the strength of the crust increases during basin inversion, so that locking of earlier inverted basins which may be incorporated within the foreland thrust belt can control and enhance the stress transmission and the progressive propagation of far-field compressional deformation farther into the plate interiors [*Ziegler et al.*, 1995].

6.2. Basement Fault Reactivation and Crustal Detachment in Forelands of Orogens

[46] The need for crustal detachments arises from the geometric and kinematic problem of accommodating crustal horizontal displacements and shortening. In the Alpine orogenic crustal wedge, such detachments allowed upper crustal stacking and accommodated related displacements/shortening of several tens or several hundreds of kilo-

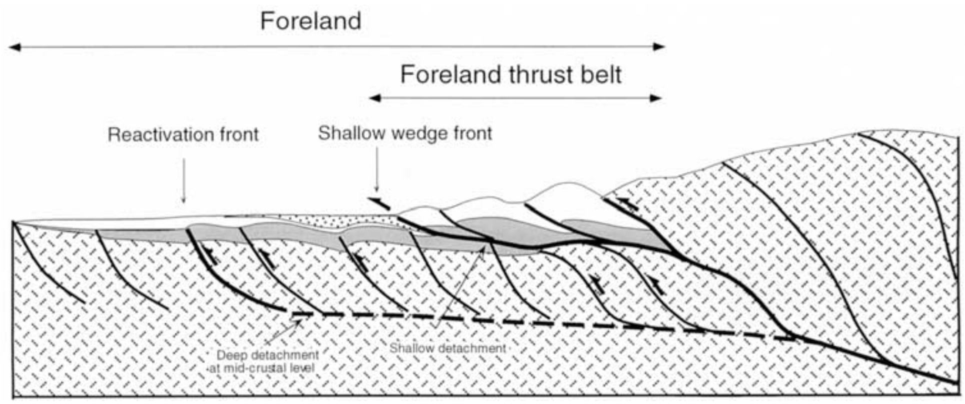


Figure 11. Schematic model of superimposed shallow and deep detachment tectonics in forelands of orogens.

meters [e.g., *Pfiffner and Hitz*, 1997]. The extension of such crustal detachment beneath the foreland, although difficult to support directly, is required for accommodating and propagating shortening far away from the orogenic front without involving the entire lithosphere with the same amount of shortening. In the outermost Alpine foreland, a detachment in the basement is required to accommodate inversion of Permo-Carboniferous grabens underlying the Mesozoic cover (Figure 7b) [*Philippe*, 1995; *Pfiffner et al.*, 1997b]. A similar situation is met in the Taiwan foreland where a detachment at 10–15 km depth probably accommodated offshore basin inversion and reversal of preexisting normal faults. Even though they remain much lower than within the orogen, cumulated displacements in forelands of orogens may reach several tens of kilometers, as estimated for instance in the North Sea through basin inversion [*Ziegler et al.*, 1995]. Presumably only a minor part of horizontal displacements can be achieved by pressure solution or oblique slip movements (i.e., shortening partly accommodated out of the plane of the section considered). In addition, the propagation and accommodation of displacements over large distances from the plate boundary, e.g., at more than 1000 km away from the Alpine front [*Ziegler*, 1987] cannot be achieved by the only involvement of a several kilometers thick detached cover, and requires that a significant thickness of crustal material be involved in shortening. This is supported by crustal seismicity which clearly argues in favor of present-day basement deformation in the Alpine or Taiwanese forelands. Balancing foreland deformation therefore requires a crustal detachment which at least partially decouples the uppermost crustal levels from the deeper lithosphere. Otherwise, it implies that deformation is distributed within the entire lithosphere which therefore undergoes the same amount of shortening than the upper crust. Although this cannot be definitely ruled out during the onset of shortening, since lithospheric folding may occur as the early response to compression [*Cloetingh et al.*, 1999], it is likely that during the evolution of the collision belt, mechanical weaknesses like inherited planar/listric faults or the brittle-ductile transition are activated, and that crystalline thrust sheets detach above such a midcrustal weak zone, as suggested by *Hatcher and Williams* [1986] and *Hatcher and Hooper* [1992]. Recent geodynamic modeling of collision tectonics [*Pfiffner et al.*, 2000] also emphasize the effect of crustal heterogeneities in promoting basement nappes formation: Results show that midcrustal weak inclusions within the underthrust lithosphere cause limited decoupling at the midcrust and localize proward shear. In most models of structural inversion in forelands, the midcrustal weak zone above which basement high-angle thrusting occurs is inherited from a previous midcrustal extensional flat detachment [e.g., *Cristallini and Ramos*, 2000; *Marshak et al.*, 2000]. Duplexes and ramps form at shallower crustal levels where the fault at the base of the basement thrust sheet can no longer propagate along the brittle-ductile transition, so that it becomes mechanically easier to ramp into the upper crust. Some works alternatively suggest that detachment of crystalline thrust sheets may occur by reactivation of moderately to steeply dipping preexisting basement faults that continue into the middle and possibly the lower crust (Urals [*Brown et al.*, 1999]), without requiring any deep detachment [*Mosar*, 1999]. This process could be important in incorporating basement thrust sheets into the thrust wedge when it is still in the early stage of its strain and thermal evolution so that detachment along the brittle-ductile transition is less (not yet) efficient.

[47] Figure 11 illustrates a geometric and kinematic model of imbricate thrust wedges, which considers that compressional deformations in forelands of orogens are accommodated by multiple, shallow, and deep detachments. The deep detachment is probably localized by crustal mechanical weakness or rheological transition and extends beneath the foreland. The shallower cover wedge, limited by ductile décollement horizon at the base of the detached cover, is the critical wedge in the sense of *Davis et al.* [1983]. This wedge belongs to a larger wedge, in which both cover and basement are involved in shortening; at its base, the shallow and deep detachments connect (Figure 11). This model has further led to discuss the actual significance of what is commonly called orogenic or deformation front, leading to clearly identify and distinguish the front of the shallow thrust wedge, the reactivation front (the outermost inverted structure), and the deformation front (the outermost microstructures related

to orogenic stresses) [*Lacombe and Mouthereau*, 1999; *Mouthereau et al.*, 2002].

[48] Since the activation of a midcrustal detachment requires a strong coupling at the crustal scale between the orogen and the foreland, the geometry of the colliding margin may have a large influence on such a coupling. In the case of Taiwan, structural inversion predominantly occurs on the southern edges of the Kuanyin and Peikang basement Highs (Figure 3b). Because presumably the rate of relative convergence between the Philippine Sea Plate and the Chinese margin did not significantly change, the most important parameter controlling the coupling at crustal level of the growing orogenic wedge with the Chinese foreland and the activation of a deep detachment is the irregular geometry of the obliquely collided margin.

6.3. Timing of Shallow and Deep Detachment Tectonics

[49] Relative timing of shallow and deep detachments in orogenic forelands needs careful consideration since it reflects the sequence of deformation at the front of the orogen. It is of considerable importance for addressing the questions of whether shortening in the basement occurred first and is transmitted to the cover, or cover detached first because of low-friction basal horizons, and deep-seated thrusting occurred second for the entire orogenic wedge to reach a new state of equilibrium.

[50] In NW Taiwan, the transpressional reactivation of normal faults in the outer Foothills and the coeval activation of the inferred deep-seated detachment predated the activation of shallow detachments and related low-angle thrusting in the inner Foothills. This suggests that the deep detachment was presumably activated first and accommodated earlier oblique inversion, and remains active after the emplacement of shallow low-angle thrust sheets as suggested by seismicity (Figure 4). This chronology is that which is expected since favorably oriented preexisting basement discontinuities constitute weakness zones available for reactivation. In central Taiwan, the Chi-Chi earthquake sequence indicates that the deep detachment and related out-of-sequence basement thrusting may occur coeval with frontal shallow thrusting. In the Pyrenean and western Alpine foreland thrust belts, the latest tectonic stage corresponds to the out-of-sequence activity of basement thrusts which presumably branch off a deep-seated detachment and postdate movement along the regional shallow detachment. However, basement uplift also occurred earlier, slightly before and coeval with emplacement of shallower nappes (see section 4). During the early stages of orogen development basement thrusting may occur in the inner parts of the orogen before the deep detachment be activated if the strain and thermal evolution of the orogen are such that detachment along the brittle-ductile transition is not yet efficient.

[51] The relative timing of deep and shallow detachment tectonics in forelands thus remains poorly constrained. No clear chronological rule can be drawn from the studied examples. Basin inversion in the far foreland, which requires the activation of a deep detachment, may occur early in the history of the foreland thrust belt. Basement thrust sheets develop in the orogen and propagate along the deep detachment (at the brittle-ductile transition). Beneath the foreland, the basal thrust ramps into the upper crust and if ductile horizons are available, into the platform sequence. Basement thrusting also occurs out-of-sequence after the emplacement of shallow thrust sheets. Mechanical boundary conditions (occurrence of a ductile evaporitic layer in the cover or the presence of preexisting weaknesses like underlying Permo-Carboniferous basins in the case of the Alpine foreland, presence of preorogenic basins in the case of Taiwan) seem to strongly control the occurrence and relative timing of thin-skinned or thick-skinned tectonics. Despite the complex interactions in space and time between both tectonic styles in the Taiwanese, Alpine, and Pyrenean forelands, the present study clearly emphasizes the overall importance of basement thrusting in the structural evolution of forelands of collisional orogens.

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