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Lutao-Lanyu Domain

The bathymetry south of Taiwan is dominated by a prominent forearc ridge and forearc basin. This antiform/synform pair, the Hengchun Ridge and the Luzon Trough, strike roughly south-southeast, parallel to the trench and arc (Figure 1a). Bathymetric relief across these features is about 3 km, and together they span a forearc that is about 150 km wide. The Hengchun Ridge is the southward, submarine extension of the Hengchun Peninsula of southern Taiwan [e.g., *Huang et al.*, 1992; *Lundberg et al.*, 1997]. This ridge is mantled mainly by middle- to late-Miocene deep marine turbidite deposits, similar to the predominant rocks of the Hengchun Peninsula [*Chen et al.*, 1985; *Sung and Wang*, 1985, 1986; *Huang et al.*, 1992]. The Luzon Trough, between the ridge and the volcanic arc, does not extend onshore. Instead, it truncates against the southeastern coast of Taiwan, where volcanic arc rocks are first juxtaposed against the metamorphic rocks that form the basement core of the forearc. Thus, the oceanic lithosphere of the Luzon Trough must have been largely consumed in the course of southward progression of orogeny in Taiwan [e.g., *Chemenda et al.*, 1997].

Since most of the Lutao-Lanyu Domain is beneath the sea, bathymetric data play a key role in interpreting its active tectonics. Long ago, *M.-P. Chen et al.* [1988] recognized that the submarine topography of the northern sector of the North Luzon Trough defines three north-south structural features. From east to west, these are the Taitung Trough, the Huatung Ridge, and the Southern Longitudinal Trough (Figure 4). More recently, higher-quality bathymetric and other data have enabled more-refined structural interpretations [e.g., *Liu et al.*, 1998; *Malavieille et al.*, 2002].

The southern limit of the ridge and two troughs, at about 21.5˚ N, may demarcate the southern front of forearc shortening [e.g., *Huang et al.*, 1992, 1997, 2000]. The east-dipping Wadati-Benioff zone of the Manila trench, however, extends under these structures as far north as Taitung [*Huang*, 1997; *Kao et al.*, 2000]. This indicates that seismic deformation of subducting Eurasian oceanic lithosphere continues beneath the entire Lutao-Lanyu Domain but does not extend farther north (dashed black line in Figure 1a).

A closer examination of the structures within the Lutao-Lanyu Domain enables a better definition of the active structures of the incipient collision (Figure 4). The Huatung Ridge is probably an anticline or an anticlinorium. Seismic reflection lines show, in fact, that sediments on the western flank of the ridge at about 22˚20' latitude tilt westward [*Huang et al.*, 1992]. The along-strike discontinuity in the ridge suggests that the magnitude of shortening across the structure varies along strike. Bathymetry and seismic reflection profiles clearly show a west-dipping thrust fault on the east flank of the ridge from about 21˚40' to 22˚20'N [*Lundberg et al.*, 1997; *Chang et al.*, 2001; *Malavieille et al.*, 2002]. Kilometer-scale wrinkles in topography west of Lanyu Island suggest many active secondary folds exist in the hanging wall block of this thrust fault. In the vicinity of 22˚N, a clear scarp exists on the west side of the ridge as well. Because of its lesser prominence, we interpret this to be a southeast-dipping back-thrust that rises off the principal west-dipping structure. This back-thrust also appears in seismic reflection profiles [e.g., *Chang et al.*, 2001]. North of about 22˚20'N, however, no scarps are obvious on either side of the ridge, so either the underlying thrust faults are blind or currently inactive.

These observations and the basic symmetry of the Huatung Ridge suggest that it is the submarine expression of a fault-bend fold above a variably west-dipping thrust fault. Growth of the fold has been incremental, as evidenced by the presence of an antecedent meandering submarine channel cutting across the feature about halfway between the latitudes of Lanyu and Lutao islands (Figure 4).

The Huatung Ridge is approximately collinear with the Coastal Range to the north. Thus the temptation is to seek structural analogies with the better-studied Coastal Range. Only part of the range is, however, analogous to the Huatung Ridge. The Coastal Range consists of three major types of geologic materials [e.g., *W.-S. Chen*, 1988; *Ho*, 1988]: Mio-Pliocene arc volcanic rocks, associated turbidite and reef sediments, and a mélange, the Lichi Formation. Clearly, we should not expect to find *in situ* arc volcanic rocks in the Huatung Ridge, except perhaps for debris sloughed off the volcanic islands. If the ridge has risen in the middle of the forearc basin, we would expect its carapace to consist predominantly of distal and proximal turbidite deposits derived from sediments exiting the southern end of the Longitudinal Valley or sweeping down the flanks of the Hengchun Ridge or the volcanic arc. Any sediment that has been highly sheared by the west-dipping thrust fault east of the ridge might be analogous to the Lichi Formation of the Coastal Range [e.g., *Biq*, 1971; *Karig*, 1973; *Teng*, 1981; *Teng and Lo*, 1985; *Chen and Wang*, 1988; *Hsü*, 1988; *Chen*, 1991; *Huang et al.*, 1992; *Chang et al.*, 2000, 2001]. Alternatively, Lichi-like mélange might also form by slump deposits associated with collapse of sediments on the steeper slopes of the ridge [e.g., *Ernst*, 1977; *Page and Suppe*, 1981; *Barrier and Muller*, 1984; *Barrier and Angelier*, 1986].

Small scarps on alluvial fans southwest of Taitung and the very linear mountain front extending south about 45 km along the coast from these small scarps suggest the presence of an active structure there (Figure 4). However, there may not be a major west-dipping thrust fault along the eastern flank of the Hengchun Ridge, since seismic reflection profiles offshore show that the eastern flank of the ridge extends eastward under the Southern Longitudinal Trough as an almost undeformed east-dipping plane [e.g., *Lundberg et al.*, 1997]. Therefore, instead of over-riding the forearc basin along a thrust fault, rocks of the Hengchun Peninsula are likely to be rising above a ramp of the thrust fault that crops out along the eastern flank of the Huatung Ridge.

High rates of uplift of the volcanic islands suggest that the volcanic arc is also involved substantially in the incipient collision of the Lutao-Lanyu Domain. Late Pleistocene rates of uplift of Lutao and Lanyu are about 3.4 mm/yr [*Chen and Liu*, 1992; *Chen*, 1993]. However, only one fault is obvious in the bathymetry: A 20-km long linear scarp west of Lutao (Figure 4) suggests that an east-dipping fault crops out there and dips under the island. If we assume a 30˚ dip for this fault, the uplift rates of Lutao and Lanyu would translate into a component of horizontal shortening perpendicular to strike of about 6 mm/yr.

In summary, bathymetric data supplemented by seismic reflection lines suggest that a major thrust fault dips west under at least the southern half of the Huatung Ridge. If an active thrust fault underlies the northern half of the ridge, it would be blind. Rapid uplift of the volcanic arc suggests the arc is also involved in the early stages of collision. A short scarp on the western side of Lutao island suggests that an east-dipping fault dips under the island.

Recent geodetic measurements by GPS reveal that Lutao and Lanyu are converging northwestward toward the Hengchun Peninsula at about 40 mm/yr [*Yu et al.*, 1997, 1999]. We propose that slip on the major west-dipping thrust fault, on the east flank of the Huatung Ridge, accommodates most of this strain. The fact that the GPS vectors are oblique to the strikes of these structures indicates that they must have a large component of left-lateral slip.

The high uplift rates of Lutao and Lanyu islands suggest that thrust faults beneath them also are contributing significantly to the shortening. The high late Pleistocene uplift rates of these islands suggest that the component of horizontal shortening on the underlying thrusts is about 6 mm/yr, perpendicular to strike. If we resolve convergence parallel to the GPS vector, the full horizontal component of slip would be about 8 mm/yr, about 20% of the total geodetically determined shortening across the domain.

At present, we cannot discriminate whether or not slip on any of these faults is seismic or aseismic, because the aperture of the geodetic measurements is many tens of kilometers. Regardless, the strains measured geodetically across this domain constitute about half of the total convergence rate between the Philippine Sea plate and the Eurasian plate [*Lundberg et al.*, 1997; *Malavieille et al.*, 2002].

Taitung Domain

The northern edge of the Lutao-Lanyu Domain is abrupt. The Southern Longitudinal Trough, the Taitung Trough and the Huatung Ridge do not continue northward beyond the latitude of Taitung city (Figure 4). The submarine ridge of the volcanic arc disappears only a little farther north. The topography and bathymetry to the north is utterly different. Figure 5 shows that the Taitung Domain is characterized by a long, narrow valley, between the Central Range and the arc-volcanic rocks of the Coastal Range, and a precipitous offshore escarpment. The active structures that have created these features are not contiguous with those of the Lutao-Lanyu Domain. In fact, the sense of vergence of the dominant structure changes abruptly across the boundary between the two domains.

The dominant active structure of the Taitung Domain is the Longitudinal Valley fault. This fault dips steeply eastward beneath the Coastal Range and locally is slipping aseismically at the surface at about 30 mm/yr. The fault has nearly equal components of sinistral and thrust motion. A second thrust fault lies just to the west, along the flank of the Central Range. This fault dips west and has both blind and emergent sections. A third thrust fault crops out on the deep sea floor about 40 km east of the coast. Accommodation structures must lie between these three principal structures of the Taitung Domain and the two principal structures of the Lutao-Lanyu Domain, but these are not readily apparent in the bathymetry.

The onshore portion of the Taitung Domain comprises three geologically distinct strips. On the east is the Coastal Range, an assemblage of young (Miocene through early Pliocene) volcanic arc rocks and associated turbidite deposits, mélange, and fringing-reef limestones [*W.-S. Chen*, 1988; *Ho*, 1988]. These rocks are similar to rocks of the forearc basin and island arc immediately to the south. Thus, the Coastal Range rocks appear to represent a highly shortened forearc basin and volcanic arc [e.g., *Chang et al.*, 2001]. To the west, the eastern flank of the Central Range is composed of Mesozoic to Paleogene schist and slate [*Ho*, 1988]. The contrast in the constitution between the two ranges demonstrates that the intervening long, linear Longitudinal Valley is a major tectonic suture [e.g., *York*, 1976; *Teng*, 1990]. Coarse-grained late Quaternary clastic fluvial sediments fill the valley. The thickness of these sediments is unknown, but could be more than a kilometer [e.g., *Chen et al.*, 1974; *Chen*, 1976].

The Longitudinal Valley fault (Figure 5) is characterized by high rates of sinistral reverse motion. Previously, identification of this fault in the field depended on scarce observations of the fault in outcrop [e.g., *Hsu*, 1956, 1976; *Wang*, 1976; *York*, 1976; *Barrier et al.*, 1982; *Chu and Yu*, 1997]. However, the geomorphic evidence of the fault is very clear along most of the Taitung Domain, although they are rather complex. Discontinuous scarps several meters high are common along the range front. As is common for thrust-fault scarps, these are lobate and irregular in plan view, and along most of the valley are accompanied by secondary anticlines and synclines in the hanging-wall block. Occasionally, scarps characterized by linear traces and other geomorphic features more typical of strike-slip structures also occur near the thrust fault scarps. In most places, the strike-slip structures lie east of the thrust fault scarps in the hanging-wall block of the main fault. Rarely, as near Yuli and east of the Wuhe Tableland (Figure 5), there are strike-slip faults lying west of the thrust fault scarp. We are unsure whether or not these strike-slip structures dip eastward at depth and are, thus, integral parts of an obliquely slipping Longitudinal Valley fault zone. They might, instead, represent separate, steeply dipping, discontinuous strike-slip faults, within the sediments of the valley.

A clear example of slip partitioning between strike-slip and dip-slip faulting occurs in the southern 20 km of the Taitung Domain [*J.-C. Lee et al.*, 1998; *Hu et al.*, 2001a; *Shyu et al.*, 2002]. Between Taitung and Rueiyuan, the fault on the eastern side of the valley is clearly left-lateral strike-slip, as evidenced by left-lateral geodetically measured creep of about 25 mm/yr, across the Taitung Bridge 3 km north of Taitung [*Yu et al.*, 1992] and linear scarps with left-lateral offset just north of Shanli [*Shyu et al.*, 2002] (Figure 5). Thrust fault morphology, on the other hand, is clear several kilometers farther west, along the western edge of a low hill referred to as both the Peinanshan and the "Foot," because of its peculiar shape (Figure 2b). This thrust fault has been called the Luyeh fault, and has progressively deformed fluvial terraces of several ages on the Foot [*Lin*, 1957; *Hsu*, 1976; *York*, 1976; *Shih et al.*, 1983a, 1984a; *Yang*, 1986; *Chu and Yu*, 1997; *Shyu et al.*, 2002]. This fault dies out gradually to the north near Rueiyuan, where GPS measurements suggest that the sinistral and thrust components partitioned onto the two separate faults rejoin along the solitary fault trace on the eastern side of the valley [*Yu and Kuo*, 2001].

Geodetic measurements confirm that the Longitudinal Valley fault is active and slipping at a high rate. Across a segment of the Longitudinal Valley fault zone near Chihshang, creepmeters show that it is creeping obliquely at a high rate [*Angelier et al.*, 1997; *Lee et al.*, 2001a]. Field measurements of deformed man-made structures indicate that the rate of horizontal shortening is about 22 mm/yr directed 323˚ [*Angelier et al.*, 1997], and leveling confirms a vertical separation rate of up to 20-24 mm/yr [*Yu and Liu*, 1989; *Yu and Kuo*, 2001]. The ratio of vertical to horizontal motion suggests that the dip of the fault in the shallow subsurface is about 45˚ eastward. However, rates of uplift of monuments along the coastline, many km east of the fault trace, are smaller than at the fault trace [*Liu and Yu*, 1990]. Moreover, uplifted mid-Holocene shorelines along the coast yield rates of only about 10 mm/yr [*Hsieh et al.*, 2001]. The fact that these rates are much lower than the rates of creep across the trace of the Longitudinal Valley fault suggests a listric geometry for the fault. Aftershocks of two moderate earthquakes in 1995 and 2003, in fact, suggest that the dip of the fault plane decreases with depth [*Chen and Rau*, 2002; *R.-J. Rau*, unpublished data].

Rupture of the Longitudinal Valley fault in the Taitung Domain produced two large earthquakes, three minutes apart, on 25 November 1951 — a M7.0 earthquake in the north, near Yuli, and a M6.2 earthquake in the south, near Chihshang [*Cheng et al.*, 1996]. The ruptures were not well documented, but there were sparse observations of fault rupture between about Chihshang and a point about 40 km north of Rueisuei [*Hsu*, 1962; *Cheng et al.*, 1996; *Shyu et al.*, 2005a] (Figure 5). The observations indicate both left-lateral and dip slip. A M6.8 earthquake in 1972 centered near Rueisuei may also have been produced by rupture of the Longitudinal Valley fault. The fault-plane solution of this earthquake indicates reverse faulting on a southeast dipping structure beneath the Coastal Range [*Chan*, 1985].

East-facing scarps on the seafloor about 50 km east of the Longitudinal Valley are evidence of active thrust faulting there. *Malavieille et al.* [2002] suggest that this fault is the principal fault of the Coastal Range block, and that the Longitudinal Valley fault is a back thrust that rises from it. We question the proposal that the Longitudinal Valley fault plays a subsidiary role, given its very high rate of slip. Such high rates suggest that the Longitudinal Valley fault is, in fact, the principal fault, and that the fault on the sea floor is subsidiary, perhaps only serving to scoot the accreting volcanic arc a few km or tens of km out over the Philippine Sea plate.

The existence of a reverse fault dipping westward beneath the eastern flank of the Central Range has long been suspected. *Biq* [1965], in fact, named this structure the Central Range fault. Several independent lines of evidence suggest its existence. First, the exposed slates of the Central Range formed at depths far below the land surface and, thus, must have been uplifted many kilometers. Second, the range's eastern flank is notably straighter than it is in the Hualien Domain to the north. Third, fluvial terraces are perched tens to hundreds of meters above modern streambeds along the eastern range front of the Central Range from at least the Wuhe Tableland to Fuli, and along the Luyeh River. Fourth, a leveling line that extends 16 km into the Central Range near Chihshang shows tilt of the easternmost 16 km of the range at about 0.8 μradian/year [*Liu and Yu*, 1990]. Thus a point 16 km into the range is rising at about 13 mm/yr relative to the Longitudinal Valley. Similar eastward tilting of the eastern Central Range is also occurring west of the Yuli area [*Yu and Kuo*, 2001]. Fifth, a seismic-refraction survey revealed that at the western edge of the Longitudinal Valley, a high-speed (~6.2 km/s) unit exists in the shallow subsurface, in direct contact with a low-speed (3.2-4.3 km/s) unit to the east. This contact may be the Central Range fault [*Chen et al.*, 1974; *Chen*, 1976]. Furthermore, recent analysis of seismicity may support the presence of a west-dipping structure beneath the eastern Central Range [e.g., *Carena et al.*, 2001].

The thrust fault that could be producing these phenomena appears to be partly blind. Along the southern half of the Taitung Domain, approximately south of Fuli, the eastern Central Range mountain front displays no small landforms indicative of an emergent young thrust fault. However, in the northern half of the domain, a west dipping thrust fault probably reaches the surface (Figure 5). For example, the fault is likely to crop out along the eastern edge of the Wuhe Tableland. The tableland is so-called because of its gently sloping surface more than 100 meters above the surrounding valley floor. The surface is a set of gently folded fluvial terraces. A blind back-thrust, or more likely, a monoclinal hinge, may delineate the tableland's western edge. Southward, numerous uplifted fluvial terraces can be found along the eastern flank of the Central Range, especially near Yuli. Small scarps at several localities along the base of these terraces indicate the fault is emergent there.

The northern boundary of the Taitung Domain is defined by a major change in the manifestation of active thrust faulting and geodetically measured strain, at about the latitude of the Ryukyu trench (Figure 3).

In summary, the principal structures of the Taitung Domain are accommodating about 40 mm/yr of obliquely convergent shortening. Most of this is currently being taken up across the rapidly creeping (but occasionally seismic) Longitudinal Valley fault, which dips east beneath the Coastal Range. A subsidiary west-dipping reverse fault crops out about 40 km offshore, on the seafloor. The Central Range thrust, which dips under the eastern flank of the Central Range, also appears to be active.

Hualien Domain

In the Hualien Domain, the Luzon volcanic arc has almost completed its accretion to the Eurasian continental shelf. Geodetic measurements suggest that convergence is occurring at only about 5 mm/yr, far slower than the 40 mm/yr of the Taitung Domain [*Yu and Kuo*, 2001]. As to the south, the dominant structure is the Longitudinal Valley fault. Geomorphic data suggest that in this domain, however, the fault is experiencing a higher ratio of sinistral to dip slip.

At first glance, the northern half of the Longitudinal Valley does not appear to be all that different from the southern half (Figure 3). As to the south, the width of the valley is about 5 km (Figure 6) and it is filled with coarse-grained clastic sediment derived predominantly from the Central Range to the west.

Despite these similarities, the Hualien Domain is fundamentally distinct from its neighbor to the south. The principal differences are that 1) the west-dipping Central Range thrust appears to be either inactive or much less active than it is to the south, 2) the east-dipping Longitudinal Valley fault has a higher ratio of sinistral to dip slip, and 3) rather than a west-dipping thrust fault offshore, the edge of the Ryukyu deformation front sits offshore.

Along the entire 65 km onland length of the Hualien Domain, the Longitudinal Valley fault crops out at the base of the western edge of the Coastal Range. Along much of that length, the fault has separate strike-slip and dip-slip strands. The strike-slip strands crop out up to a km or so east of the dip-slip strands. Geomorphic features characteristic of strike-slip faults, such as offset streams and linear ridges, are common along these traces. Geomorphic features along the dip-slip traces of the fault are those commonly seen along thrust faults – hummocky topography and the rollover of surfaces in the hanging wall block. The strike-slip traces of the fault are much more prominent and continuous than they are in the Taitung Domain. The thrust fault scarps seldom rise more than 20 meters above the valley floor. The dominance of left-lateral movement on the northern Longitudinal Valley fault system near Hualien is also clear in both early triangulation data [*C.-Y. Chen*, 1974] and in more recent leveling data [*Yang*, 1999]. The average elevation of the Coastal Range is also lower in the Hualien Domain. The highest peaks in the northern Coastal Range are all less than 1000 m above sea level, whereas many peaks to the south are much higher, with the highest peak at 1682 meters.

Seismic rupture of a section of the Longitudinal Valley fault provides additional support for the predominance of strike-slip over dip-slip. Rupture of the Meilun strand, which runs along the western edge of the Meilun Tableland near Hualien City, produced a M7.3 earthquake in October 1951 [*Yang*, 1953; *Hsu*, 1962]. Vertical offsets across the rupture were about 1.2 meters, and left-lateral offsets were about 2 meters [*Hsu*, 1955; *Lin*, 1957; *Bonilla*, 1975, 1977; *Yu*, 1997]. Minor scarps on the tableland are also likely to reflect active faults, and most of them may also have slipped during the 1951 earthquake [*Taiwan Weather Bureau*, 1951; *Yu*, 1994, 1997].

A submarine ridge extends northeastward approximately 10 km from the Meilun Tableland (Figure 6). The steep and linear western flank of the ridge appears to be the continuation of the Meilun fault strand [*Tsao*, 1975]. Further northeast is another submarine ridge, which trends nearly E-W and seems to be unconnected to any onland topography. The abrupt and steep scarp at its western edge is likely to be a further extension of the Meilun fault strand. Moreover, the northern flank of this ridge is another steep scarp, which separates the ridge from the much lower and smoother Hoping Basin. These scarps appear to represent the initiation of the eastward peeling off of the Coastal Range from the Taiwan orogen. As one would expect in a region of transition between suturing and unsuturing, many minor structures are present in this area. We will discuss this transition zone in conjunction with discussion of the Ryukyu Domain.

We find scant evidence for activity of the Central Range thrust within the Hualien Domain. The rarity of high lateritic fluvial terraces in the river drainages of the eastern flank of the range [*Yang*, 1986; *Chang et al.*, 1992] suggests lower uplift rates than to the south. Moreover, the sinuosity of the eastern flank of the Central Range indicates that rates of alluvial deposition outpace rates of uplift of the range. The sinuosity is due to the fact that the apices of large alluvial fans delivering debris to the valley extend upstream many km into the Central Range [*Chang et al.*, 1994]. This would not be the case if uplift rates exceeded rates of alluviation. One might argue that alluvial sedimentation rates are greater in the north than in the south. This is implausible, however, since the average drainage basin area, the average slope of the rivers, the bedrock type and the weathering patterns are very similar in the two domains. Thus it appears that the Longitudinal Valley fault is the only major active structure of the onshore part of the Hualien Domain.

The southern and northern boundaries of the Hualien Domain are well defined. The southern boundary appears to be a 20-km-wide transition zone between Rueisuei and Kuangfu (Figure 6). At Rueisuei, the northernmost clear evidence for activity of the Central Range thrust is present. There the traces of the Central Range thrust and Longitudinal Valley thrust come within 300 meters or less of each other. Immediately north of this intersection, the trace of the Longitudinal Valley thrust fault deviates markedly to the west. Between Rueisuei and Kuangfu, the geomorphology of the fault trace suggests that thrusting is dominant over sinistral slip, as in the Taitung Domain to the south. But we find no indication of activity of the Central Range thrust along this portion of the valley. Thus, this section of the valley is transitional between the two domains. Recent GPS measurements support this conclusion. They show that the transition from highly oblique convergence to predominant sinistral motion spans a ~15-km length of the valley, near Kuangfu. South of Kuangfu, the convergence rate normal to the strike of the Longitudinal Valley fault is as high as 30 mm/yr [*Yu et al.*, 1997, 1999]. North of Kuangfu, however, the convergence rate is dramatically lower—just 5 mm/yr. The left-lateral component of the GPS vectors does not change appreciably across this region [*Yu and Kuo*, 2001]. Finally, the limit of rupture of the November 1951 rupture may signal the transition between domains. The northern limit of reported rupture along the fault in 1951 is near Kuangfu [*Hsu*, 1962].

This striking difference in the component of convergence across the transition from Taitung to Hualien Domains has inspired several researchers [e.g., *Angelier et al.*, 1997; *Hu et al.*, 2001b] to propose a hypothetical active fault, which cuts through the Coastal Range and extends offshore at a point between Fengpin and Hualien. Right-lateral and dip slip on this hypothetical fault would serve to accommodate the contrast in magnitude of convergence between the Taitung and Hualien Domains. However, geomorphic and structural data enable us to argue against the existence of such a fault. The dramatic change in convergence rate across the domain boundary appears to reflect an abrupt northward slowing of the accretional process of the collision, but a discrete accommodation structure is not apparent at the surface. Instead, the change in convergence rates is more probably accommodated by clockwise rotation of a section of the Coastal Range across the transition zone. A denser GPS array in this part of the Coastal Range would enable a test of this idea.

The eastern and northern boundaries of the Hualien Domain are the edge of the Ryukyu subduction system. North of Hualien, the Coastal Range and Longitudinal Valley drop to sea level, and the sea laps against a precipitous eastern flank of the Central Range (Figure 3). This is the southern edge of the extension associated with the Ryukyu subduction system and the tearing apart of the forearc/volcanic arc suture. To the east is the deformation front of the Ryukyu subduction zone. We will now turn our attention to these extensional and subduction elements of the Ryukyu Domain.

Ryukyu Domain

The Ryukyu Domain encompasses that part of the Ryukyu subduction megathrust nearest Taiwan that is capable of producing large subduction earthquakes. We have drawn the boundaries of the domain with the intention of encompassing all active structures between the trench and the southern side of the back-arc Okinawa Trough. The dominant structure in this domain is, of course, the subduction interface, but accommodation structures on the western edge of the domain are also important. In this section, we discuss the principal structure first, and then turn to the transitional structures on the western edge of the domain.

Isobaths drawn on the top of the Wadati-Benioff zone show that the dip of the Ryukyu subduction interface averages about 40˚ to a depth of about 50 km and steepens downdip from there to more than 55˚. The Wadati-Benioff zone defines the subducting slab to a depth of greater than 250 km [*Kao et al.*, 1998; *Kao and Rau*, 1999]. The interface is approximately 120-140 km beneath Taipei [*Kao et al.*, 1998]. Along the western portion of the subduction zone, the isobaths shallower than 50 km curve gently westward. This indicates that the westernmost part of the interface bends northward and transforms from a gently dipping surface to a nearly vertical plane [*Kao et al.*, 1998]. The termination of seismicity at the western edge of the Wadati-Benioff zone is abrupt. It trends nearly north-south, from the edge of the Ryukyu Domain to the western edge of the Taipei Domain (dashed black line in Figure 1a).

Secondary thrust faults in the hanging-wall block are also apparent. Immediately north of the Ryukyu trench is the Yaeyama forearc ridge [*Liu et al.*, 1998] (Figures 3 and

7), a 40-km-wide feature that consists of a series of E-W striking anticlinal ridges and synclinal valleys. These are likely the manifestation of trench-orthogonal shortening of an accretionary prism above the subduction interface [e.g., *Lallemand et al.*, 1999; *Font et al.*, 2001]. The folds may be the result of slip on thrust faults that splay upward from the subduction interface. Also, two major right-lateral faults in the northern portion of the ridge have been proposed to accommodate the oblique subduction direction [e.g., *Lallemand et al.*, 1999; *Chemenda et al.*, 2000; *Font et al.*, 2001].

North of this fold-and-thrust belt of the accretionary prism is an irregular 40-km-forearc trough that includes the Hoping and Nanao forearc basins [*Liu et al.*, 1998] (Figures 3 and 7). The steep escarpments that flank the trough suggest the presence of active structures. However, seismic reflection profiles across the northern flank indicate that the sediments of the basin drape the flank of the trough and are not faulted [e.g., *Lallemand et al.*, 1997; *Schnürle et al.*, 1998]. The southern flank, on the other hand, is clearly faulted, with deformed and offset sediment layers in the seismic reflection profiles [e.g., *Schnürle et al.*, 1998; *Font et al.*, 2001].

North of the forearc basins is the Ryukyu arc, an ~80-km-wide submarine ridge with several small subaerial islands. We have found no clear evidence for young faulting on top the ridge, nor on the southern slope of the ridge. As mentioned earlier, the Ryukyu arc may well be constructed on the ancient forearc ridge of the Luzon arc on the Philippine Sea plate and may be the submarine, northeast extension of the Central Range [*Shyu et al.*, 2005b].

The western margin of the Ryukyu Domain is a very complex transition zone between three domains: the Hualien, Ryukyu and Ilan Domains. Thus, in this region, one may expect to find numerous minor accommodation structures. Also, bathymetry shows that the westernmost part of the Ryukyu deformation front swings abruptly northward (Figure 7). This abrupt turn of the deformation front is associated with the western edge of the Wadati-Benioff zone (Figure 1a). The linearity of the scarp along this north-trending segment suggests a substantial component of right-lateral strike-slip. This is supported by the fact that the westernmost part of the Ryukyu subduction interface bends northward and transforms into a nearly vertical plane [*Kao et al.*, 1998]. Two more right-lateral faults are present in the eastern part of the Hoping basin. These structures probably aid in the southward translation of the Ryukyu arc associated with the opening of the Okinawa Trough.

On May 15th, 2002, a moderate earthquake (M_w = 6.0) resulted from slip on a related right-lateral strike-slip fault. The epicenter of the earthquake was offshore, about 25 km north of Nanao. According to the database of the Broadband Array in Taiwan for Seismology (BATS; http://bats.earth.sinica.edu.tw/), the fault plane solutions of the main shock and its aftershocks showed right-lateral movement on an approximately N-S striking plane. This structure is properly situated and oriented to serve as one of the structures that form the western boundary of the Ryukyu subduction zone.

Other transitional structures are also present. The steep scarp at the southern edge of the Hoping basin may result from a normal fault cropping out at its base. North of Hualien, the coastline traces its course along a high, steep cliff. The cliff exposes metamorphic bedrock of the Central Range. Although the embayed river mouths near Nanao [*Hsu*, 1978] are not typical of scarps along normal faults, several short scarps farther back into the range appear to reflect minor active normal faulting. Thus, we believe the coastal cliff is the result of normal faulting. Thus, the Hoping basin appears to be a triangular shaped graben that opens to the northeast. This geometry suggests it is the product of initial post-collisional extension.

Kaoping Domain

The Kaoping Domain, the southernmost of the western domains, encompasses the final episode of subduction of oceanic lithosphere prior to collision of forearc ridge and continental margin. Its northern boundary coincides with the northern limit of the Wadati-Benioff zone of the Manila trench [e.g., *Huang et al.*, 1992; *Kao et al.*, 2000] (dashed black line in Figure 1a). Its northern boundary also coincides with the impingement of the Western Foothills and the Central Range, which we infer to reflect the elimination of oceanic lithosphere from between the edge of the Eurasian continental shelf and the forearc ridge. Unlike the more mature collisional domains farther north, the Kaoping Domain includes a rapidly subsiding basin, between the Central Range and the Western Foothills. We interpret this to be analogous to the submarine part of the Lutao-Lanyu Domain, its neighbor to the east: these are the last vestiges of oceanic lithosphere to be consumed prior to suturing (converging arrows in Figure 1a).

Both the submarine and subaerial sectors of the Kaoping Domain consist of three tracts. On the east is the forearc ridge, which continues onto land as the rapidly rising Hengchun Peninsula and Central Range (Figures 3 and 8). The peninsula and range consist of deep-marine turbidites and mélange in the south and metamorphosed slate and continental basement rocks in the north. This south-to-north progression of lithology indicates the progressive uplift and unroofing of the basement of the forearc ridge across the Kaoping Domain. On the western margin of the domain is a 50-km-wide submarine fold-and-thrust belt above the shallowest part of the Manila trench. This appears to extend onto land as the subaerial belt of deformation between Tainan and the Western Foothills. Between these eastern and western structural elements is a structurally non-descript tract of what must be marine lithosphere overlain mostly by deep-sea sediments. This extends onshore to the Pingtung Plain, where subsidence is occurring at prodigious rates.

On the western flank of the Central Range, Miocene slates are metamorphosed marine sediments, possibly deposits that draped the outer-arc ridge. The southern extension of the Central Range at the southern tip of Taiwan is the Hengchun Peninsula. Rocks of this lower ridge consist of Miocene to early Pliocene shallow to deep marine sandstones, mudstones and mélange [e.g., *Chen et al.*, 1985; *Sung and Wang*, 1985, 1986; *H.-Y. Lin*, 1998]. These sediments appear to be emerged accretionary wedge deposits [e.g., *Huang et al.*, 1992], and are very likely to be the unmetamorphosed equivalent of the slates of the Central Range.

In the west of the Kaoping Domain is the southernmost part of the Western Foothills, a strip of low, hilly terrain. Rocks of the Western Foothills are dominated by Miocene to late Pliocene sandstones and mudstones [*Ho*, 1988]. Although similar in age, these sediments are distinct from the rocks of the western Central Range in that they are mostly shallow continental shelf deposits. The deposition of some of the sediments could be in environments as shallow as coastal or even fluvial [*Yue*, 1997; *Yu*, 2001]. It is therefore likely that the Western Foothills are sediments scraped off continental shelf and incorporated into the fold and thrust belt. At the northwestern corner of the domain, incipient growth of the Western Foothills is manifest as several sub-parallel folds that interrupt the otherwise flat topography of the Chianan Coastal Plain. This is best illustrated by the low tableland around Tainan, where the deformation front appears to make landfall.

Between the Central Range and the Western Foothills lies the Pingtung Plain, a wide fluvial valley with thick Quaternary deposits. The upper hundred or so meters of sediments are coastal to estuarine sand and mud, rich in shallow-marine to lagoonal shells and foraminifera and deposited during the latest sea-level transgression [*Shyu*, 1999]. The rapid subsidence of the Pingtung Plain, especially near the coast, has caused many land-use problems in the area. Although present-day subsidence rates of up to 100 mm/yr [*Hou et al.*, 1998] appear to be due mostly to the pumping of ground water, this historical phenomenon only accounts for about 3 meters of total subsidence [*Hsu et al.*, 1998; *Chien et al.*, 1998]. An average Holocene subsidence rate of about 13 mm/yr has been determined using dates from sediment cores from the plain [*Lu et al.*, 1998; *Shyu*, 1999]. This rapid long-term subsidence rate supports the idea that an oceanic crystalline basement beneath the plain is sinking in this last stage of elimination of oceanic lithosphere between continental margin and forearc ridge.

The Chaochou fault separates the Pingtung Plain from the Central Range (Figure 8). The presence of Miocene-age rocks in the mountains on the east and a thick section of Quaternary strata in the basin to the west demonstrates that the fault has a significant component of vertical slip, up on the east [e.g., *Ho*, 1988]. The linearity of this fault suggests that it also has a significant component of strike-slip motion. This strike-slip component is also apparent in the details of morphology of the mountain front. Along the northern portion of this mountain front, a series of uplifted terrace surfaces are present. The eastern edges of these terraces are east-facing scarps, which are associated with linear ridges and troughs suggestive of strike-slip motion [*Bonilla*, 1975; *Shih and Teng*, 1983; *Shih et al.*, 1984b; *Yang*, 1986]. South of the Ailiao River (Figure 8), an abandoned linear river valley suggests the possibility that sinistral displacement could be as great as 10 kilometers. These sinistral faults appear to form a major high-angle fault zone [*Hsieh*, 1970; *Chiang*, 1971; *Yu et al.*, 1983]. The western edge of the uplifted terraces, on the other hand, is a reverse fault. Locally, it has produced scarps tens of meters high [*Hsu and Chang*, 1979; *Shih and Teng*, 1983; *Shih et al.*, 1984b; *Yang*, 1986]. Two of the major terrace surfaces are warped into anticlines. Together these strike-slip and reverse faults may represent strain partitioning of a steep east-dipping oblique-slip fault [*Chiang*, 1971; *Yang*, 1986].

Some previous mapping [e.g., *Wu*, 1978; *Ho*, 1988] inferred that the Chaochou fault continues southward along the western front of the Hengchun Peninsula and becomes the Hengchun fault [e.g., *Shih et al.*, 1985a; *Yang*, 1986] near the southern tip of the peninsula, just east of a prominent east-tilted Quaternary fluvial terrace (Figure 8). Alternatively, *Malavieille et al.* [2002] suggest that the Chaochou fault runs southward into a submarine canyon west of the Hengchun Peninsula (Figure 3). The east-tilted fluvial surface west of the town of Hengchun appears to be the result of slip on yet another east-dipping reverse fault that would crop out on the sea floor west of the coast [*Biq*, 1972]. Dating of Holocene coral terraces along the coastline reveals that the tilted surface has been rising at rates as high as 5-7 mm/yr [*Chen*, 1993; *Chen and Liu*, 1993].

The western flank of the Pingtung Plain may also represent an active fault (Figure 8). Although this low escarpment has clearly been modified by erosion of the Kaoping River, sediment cores retrieved from the basin suggest that pre-Quaternary rocks are more than 200 meters lower in the basin than they are in the hills to the west [*Wu and Tsai*, 1996; *Wu*, 1997; *Shyu*, 1999]. This indicates that a fault with at least several hundred meters of vertical slip separates the rocks of the hills from those of the basin.

Offshore to the southwest are numerous anticlinal ridges and minor thrust faults on the seafloor [e.g., *Liu et al.*, 1997]. Some of these ridges may be mud diapirs [e.g., *Huang*, 1993; *Sun and Liu*, 1993; *Yu and Lu*, 1995]. However, many of the ridges are too long and narrow to be diapirs and are more likely to represent a submarine fold-and thrust belt [e.g., *Liu*, 1993; *Yu*, 1993; *Fuh et al.*, 1994; *Lacombe et al.*, 1999; *Yu and Song*, 2000] (Figure 8). Several submarine canyons cut through a series of these ridges, clearly indicating their antecedence and the incremental growth of the folds. We interpret this submarine fold-and thrust belt to be the result of the shortening of the forearc between the continental margin and the forearc Hengchun Ridge, and the embryonic form of Taiwan's Western Foothills.

On land, NE-SW striking right-lateral faults and N-S striking folds dominate the deformation of the Western Foothills (Figure 8). The orientations of these structures indicate an overall N-S stretching and E-W shortening of the shallow crust. A series of short, NE-SW striking scarps west of the Pingtung Plain in the Western Foothills characterize the strike-slip faults. These scarps also include several previously documented faults in the suburban area north of Kaohsiung City [*Sun*, 1964; *Hsu and Chang*, 1979; *Yang*, 1986]. Locally, these scarps show right-lateral offset. Individual offsets appear to be small, but these structures form a wide belt of dextral deformation.

The crust between these right-lateral structures and the two major left-lateral structures of the Pingtung Plain must be extruding southwestward [e.g., *Lacombe et al.*, 1999, 2001; *Chan et al.*, 2001, 2002].

Also in the Western Foothills of the Kaoping Domain are a series of sub-parallel, N-S trending folds [*Meng*, 1967]. Some of these are anticlines less than 10 kilometers long. These anticlines typically form high ridges, covered by late Quaternary coralline limestone [*Ichikawa*, 1927; *Hanzawa*, 1931; *Heim and Chung*, 1962; *Sun*, 1963]. They are generally considered to result from mud diapirs on the sea floor, as illustrated by the widespread mud diapirs offshore southwestern Taiwan [e.g., *Huang*, 1993; *Liu*, 1993; *Sun and Liu*, 1993; *Yu*, 1993]. Alternatively, they may be fault-propagation folds [e.g., *Lacombe et al.*, 1997]. Many young fluvial terraces in the Western Foothills clearly show deformation across these active folds [*Hsieh*, 1999; *Hsieh and Knuepfer*, 2002]. Among these structures are the anticline-syncline pair underlying the city of Tainan [*Hsieh*, 1972; *Wu et al.*, 1992] (Figure 8). Although some of the anticlines in this group may have been produced by mud diapirism, the location of the folds just east of the deformation front suggests that they represent the initial deformation of the coastal plain. A recent SAR interferometry analysis suggests that the Tainan area is rising aseismically at a rate of ~16 mm/yr [*Fruneau et al.*, 2001]. Some have argued that this is an anthropogenic phenomenon, possibly resulting from groundwater withdrawal, since the average Holocene uplift rate of the anticline under Tainan is only about 5 mm/yr [*Chen and Liu*, 2000]. If instead it is of tectonic origin, the aseismic nature of the uplift may be an indication that motions on structures of this domain are partially aseismic, even at shallow depth.

In summary, deformation in the Kaoping Domain reflects E-W shortening and southward extrusion of sediments above a foundering oceanic lithosphere [e.g., *Lacombe et al.*, 1999, 2001]. This shortening reduces the width of the domain from more than 100 km in the south to less than 50 km in the north. With the shortening, a broad offshore fold and thrust belt in the south transforms into the nascent, narrow Western Foothills in the north, characterized by several prominent N-S trending folds and NE-SW trending right-lateral faults. Geodetic observations suggest that some of these structures may be slipping aseismically. Farther east, the rapidly-sinking Pingtung Plain is bounded by left-lateral faults that facilitate its southwestward extrusion, and is likely to be above the last piece of sundering South China Sea oceanic lithosphere. GPS observations support the conclusion that both the Chaochou fault and the unnamed fault bounding the Pingtung Plain on the west are dominated by sinistral deformations, and that the entire area is moving southwestward [*Yu et al.*, 1997]. It is noteworthy that, even so, the vectors on the west coast still require a significant right-lateral component of motion on some as-yet-unknown fault offshore, between the coast and the Penghu Islands.

Chiayi Domain

Farther north, in the Chiayi Domain, the collision of forearc ridge and continental shelf has begun. There is no intervening basin analogous to the Pingtung Plain between the Western Foothills and the Central Range (Figures 3 and 9). Instead these two ranges abut along a fault zone that displays no clear evidence of activity. Furthermore, the Kaoping Domain’s complex of small active folds and strike-slip faults within the Western Foothills does not extend northward into the Chiayi Domain. Rather, the folds of the Western Foothills there display no clear evidence for activity. Instead, the most prominent geomorphic evidence of youthful deformation exists in the low Chiayi Hills, farther west. Young lateritic fluvial terraces and active streambeds display evidence of recent folding along axes that trend nearly north-south. The geometry of these folds is broadly consistent with the presence of a shallow-dipping blind thrust fault complex beneath the Chiayi Hills and the coastal plain.

GPS geodesy reveals a pattern of horizontal strain that is quite consistent with the geomorphic pattern of late Quaternary deformation. Unlike the Kaoping Domain, deformation across the Chiayi Domain is dominated by vectors that trend northwestward, perpendicular to the strike of the active folds. The difference in magnitude of the vectors across the domain indicates NW-SE shortening at a rate of about 30 mm/yr. Historically active right-lateral strike-slip faults near the southern and northern ends of this blind thrust system are parts of the transition zones that separate the blind thrust system of the Chiayi Domain from neighboring domains to the south and north.

From east to west, there are four physiographic and stratigraphic belts comprising the Chiayi Domain: the Central Range, the Western Foothills, the Chiayi Hills and the Chianan Coastal Plain. The major rocks of the Central Range are metamorphosed Miocene marine sediments, and the rocks of the Western Foothills consist mostly of Miocene to late Pliocene sandstones and mudstones deposited on the shallow-marine continental margin. NNE-SSW-striking ridges and valleys in the Western Foothills are the geomorphic expressions of major folds or minor faults in the western fold and thrust belt of Taiwan, which are the products of fault-bend folding over an underlying décollement [*Suppe*, 1976, 1980]. These folds are likely to be the more mature equivalent of the N-S striking folds in the Kaoping Domain to the south.

The low hills immediately west of the Western Foothills are commonly referred to as the Chiayi Hills. These hills are mantled with uplifted late-Quaternary lateritic fluvial terraces and consist of Quaternary fluvial sandstones and conglomerates. Between the Chiayi Hills and the west coast is the gently sloping Chianan Coastal Plain, the most important agricultural area in Taiwan. This basin is filled with Quaternary coastal and fluvial sediments up to 1500-2000 meters thick [*Stach*, 1958; *Hsiao*, 1971, 1974; *Sun*, 1970, 1971, 1972; *Tong and Yang*, 1999; *Yeh et al.*, 1999].

Bedrock structures east of the Chiayi Hills do not exhibit evidence of recent activity. Although the Laonung fault, the boundary between the Central Range and the Western

Foothills, follows approximately a major river valley (Figure 9), detailed geomorphic features indicate that this fault is inactive, since neither the tributary valleys nor the ridgelines are disturbed by the fault. The Chukou thrust fault is the boundary between the Miocene to late Pliocene rocks of the Western Foothills and the Quaternary sandstones and conglomerates of the Chiayi Hills. For several reasons, this fault is considered by some to be the major active structure of the region [e.g., *H.-C. Chang et al.*, 1998; *Lin et al.*, 2000a]. First, this fault is contiguous with the Chelungpu fault, which ruptured in 1999. Thus some consider these two structures to be segments of the same fault zone and to be contemporaneous. Second, the Chukou fault crops out along the front of the Western Foothills, which suggests it is active. Third, a gradient of about 30 mm/yr in horizontal GPS velocities straddles this structure [*Yu et al.*, 1997; *Yu and Chen*, 1998; *Hung et al.*, 1999]. Fourth, geomorphologic and paleoseismic analyses indicate that this fault may have slipped at 38 ka [*Huang et al.*, 1994a]. Finally, there have been many earthquakes in the region [*Wu et al.*, 1979; *Yu et al.*, 1983], including the hypocenter of a M6.2 earthquake in 1998, which was very close to the down-dip projection of the Chukou fault.

None of these five observations is compelling evidence that the Chukou fault is active. In fact, geomorphic evidence suggests that this fault is inactive, or at the very least is not absorbing more than a small percentage of the 30 mm/yr of shortening that is occurring across the Chiayi Domain. First, ruptures in 1999 along the Chelungpu fault did not continue onto the Chukou fault. In fact, co-seismic and post-seismic deformations measured on scarps in the field and by GPS indicate that the deformation might continue onto a right-lateral strike-slip fault within the transition zone between the Chiayi Domain and the domain to the north [*Central Geological Survey*, 1999a; *Y.-H. Lee et al.*, 2002]. Strictly speaking, the Chelungpu thrust fault ends at the point where this strike-slip tear and the thrust fault join. Second, although the Chukou fault crops out along the western flank of the Western Foothills, it is not along the base of the mountain.

In most places the fault crops out in the middle of the western slope of the mountain front. Numerous uplifted hills and terraces are present west of the slope, in the footwall block of the Chukou fault, and these terraces have been deformed by folding [e.g., *Chung*, 1968; *Chyi and Sung*, 2000]. Third, the 30-mm/yr GPS gradient mentioned above is rather broad, and the decrease in horizontal velocities extends well into the footwall block of the Chukou fault [*Yu and Chen*, 1998; *Hung et al.*, 1999]. Fourth, after 38 ka, there is no evidence that the Chukou fault has been active. And finally, detailed three-dimensional mapping shows that the aftershocks of the 1998 earthquake are distributed on a surface deeper than the Chukou fault [*Lo*, 2001].

In contrast, evidence for youthful tectonic deformation is abundant in the Chiayi Hills and the Chianan Coastal Plain, west of the Chukou fault (Figure 9). Uplifted fluvial terraces in the Chiayi Hills west of the Chukou fault clearly indicate deformation across several folds and minor faults [e.g., *Huang*, 1996; *Chen*, 1999]. Progressively tilted lateritic surfaces indicate ongoing deformation over the past few tens of thousands of years. Across the Hsiaomei anticline, such folding even deforms the present day river bed [*Chen*, 1999; *R.-F. Chen et al.*, 2001]. Terraces along the major rivers of the Chianan Coastal Plain indicate regional uplift. Moreover, below the surface of the coastal plain, presence of minor faults and folds has been inferred in some sediment core analyses and in seismic surveys [e.g., *Tsai and Chiu*, 1976; *Sun et al.*, 1998, 1999; *Lee*, 2000].

The evidence for anticlinal deformation of lateritic terraces and modern streambeds belies the presence of a major active thrust fault below the Chiayi Hills and the Chianan Coastal Plain. This deeper structure has long been suspected [*Suppe*, 1980; *Hung*, 1996; *Hung et al.*, 1999] and is likely to be the blind décollement of the western Taiwan fold-and-thrust belt. The deformation front associated with this blind structure, just a few kilometers inland from the west coast, is aligned with the deformation front of the Kaoping Domain, to the south. Small folds of the Chiayi Domain, such as the Hsiaomei anticline, may be the result of irregularities in the underlying décollement and secondary structures above it [*Suppe*, 1976, 1980].

In the past decade, several moderate earthquakes have occurred in the Chiayi area. These include the M5.9 Tapu earthquake of 1993, M6.2 Rueyli earthquake of 1998, and the M6.4 Chiayi earthquake of 1999 [e.g., *Lo*, 2001; *R.-J. Rau*, unpublished data]. Although most of the hypocenters of these earthquakes and their aftershocks are along a shallow and flat detachment underlying the Chianan Coastal Plain, many other earthquakes also occurred along steeply-dipping planes below the detachment [*R.-J. Rau*, unpublished data]. In order to interpret these deeper seismic events, *Mouthereau et al.* [2002] have suggested the existence of another, deeper detachment beneath the Chiayi area, within the pre-Tertiary basement rocks. Alternatively, these deeper earthquakes may reflect the re-activation of pre-existing high-angle normal faults under the coastal plain sediments [e.g., *Hsiao*, 1971; *Chang et al.*, 1996; *C.-I. Lee et al.*, 2002]. A third possibility is that the earthquakes are being caused by rupture of small faults distributed broadly in the basement rocks below the décollement.

Recent GPS measurements of the velocity field within the Chiayi Domain are consistent with the neotectonic evidence for a shallow blind master thrust fault. For the period between 1993 and 1999, GPS velocity vectors for the handful of stations within the domain show a northwest decrease in convergence rates that amounts to about 30 mm/yr over the 70-km width of domain [e.g., *Yu et al.*, 1997; *Yu and Chen*, 1998; *Hung et al.*, 1999]. Aseismic slip on a 2° to 11° east-dipping décollement, down-dip from a brittle-ductile transition beneath the Western Foothills, mimics this broad gradient, if the rate on the shallow-dipping detachment is about 35 mm/yr and its depth is about 8-9 kilometers [*Hsu et al.*, 2003]. Moreover, according to their model, the up-dip terminus of aseismic slip on the décollement is about 8 km beneath the trace of the Chukou fault. This implies that the up-dip section of the décollement, west of the Chukou fault and beneath it is locked and the future source of large earthquakes [*Hsu et al.*, 2003].

The surface ruptures associated with two large destructive historical earthquakes provide important insights regarding the northern and southern boundaries of the Chiayi Domain. A M7.1 earthquake in 1906 occurred near the northern boundary of the domain. This earthquake was accompanied by right-lateral ruptures along the ENE striking Meishan fault (Figure 9). The rupture was about 13.5 kilometers in length, and its maximum dextral and vertical displacements were 2.4 meters and 1.8 meters [*Omori*, 1907; *Bonilla*, 1975, 1977; *Hsu and Chang*, 1979; *Huang et al.*, 1985]. The vertical component changed sense along the course of the rupture, as is common for ruptures dominated by lateral movement [*Biq*, 1976, 1991].

The southern few kilometers of the 1999 Chi-Chi rupture was a strike-slip fault that appears to be collinear with the 1906 Meishan rupture [*Central Geological Survey*, 1999a, b; *Lin et al.*, 2000b]. Together, these two segments formed an important boundary between two blocks moving separately during and after the 1999 Chi-Chi earthquake [*Y.-H. Lee et al.*, 2002], revealed by a GPS based co-seismic and post-seismic deformation analysis. Therefore, the Meishan fault and the terminal right-lateral fault of the 1999 rupture serve as accommodation structures between the blind thrust décollement of the Chiayi Domain and the active thrust faults of the neighboring domain to the north.

Another historic fault rupture appears to have served a similar purpose at the southern end of the Chiayi Domain. This N70˚E rupture of the right-lateral Hsinhua fault produced the M6.8 earthquake in 1946 (Figure 9). The rupture was at least 6 kilometers and perhaps more than 12 kilometers long [*Chang et al.*, 1947; *Bonilla*, 1977]. Dextral displacement of up to 2 meters prevailed, and the maximum vertical displacement reached 0.76 meters. All surface evidence of the fault has been obliterated by human activity, so the location of the fault is known solely from historical documents and shallow seismic reflection surveys [*Wang et al.*, 1994; *Shih et al.*, 1998; *Wu*, 1998]. Judging from its location, orientation and sense of slip, this structure is also likely to be an accommodation structure in the transition zone between the Kaoping and Chiayi

Domains (Figure 3).

A few other dextral-slip faults may exist within the Chiayi Domain, as evidenced from seismic profiles [e.g., *Shen and Wang*, 1999]. These may be re-activated high-angle normal faults that strike ENE to E-W in the basement rocks beneath the sediments of the Chianan Coastal Plain, [*Hsiao*, 1971, 1974; *Chow et al.*, 1988; *Chang et al.*, 1996]. Depositional patterns and differential subsidence rates across these faults [*S.-Y. Lin*, 1998; *Wu*, 1999, 2000; *Huang*, 2001] suggest that some may have been recently re-activated.

Taichung Domain

In the center of the western Taiwan orogenic belt, imbricated thrust sheets with many kilometers of accumulated slip reveal that the collision is well-established. Two sub-parallel thrust faults, the Chelungpu in the east and the Changhua in the west, are the major active components (Figure 10) of the Taichung Domain. These structures merge and die out near the southern edge of the domain and through a complex 20-km-wide transition zone in the north.

The Taichung Domain consists of four geomorphically and structurally distinct strips (Figure 10). From west to east, these are the coastal plain, the Pakua and other tablelands, the Taichung piggyback basin and the Western Foothills. The coastal plain in the west is an unfaulted, unfolded portion of the foreland basin, underlain by more than a thousand meters of thick Quaternary gravels and sands derived by erosion of the deforming terrains to the east [e.g., *Stach*, 1958]. East of this coastal strip lies a chain of uplifted and folded late-Quaternary lateritic fluvial surfaces, underlain by Quaternary coastal and fluvial sandstones and conglomerates [*Ko*, 1997]. These gently deformed Houli, Tatu, and Pakua Tablelands have risen above the surrounding plains on the back of the underlying Changhua thrust [*Chang*, 1971; *Chen*, 1978]. East of the tablelands is the Taichung foreland basin, which rides piggyback on the Changhua thrust and receives debris from the rising hills to the east. Late Quaternary fluvial sediments in this basin reach thicknesses of 3 km [*Chang*, 1971]. Abutting the Taichung basin on the east are the Western Foothills, which consist of Mio-Pliocene shallow continental shelf sediments and Plio-Pleistocene foreland basin deposits. These sedimentary rocks form the hanging-wall block of the Chelungpu thrust fault.

The 1999 rupture of the Chelungpu fault provided the critical first clue that the active Taiwan orogeny segregates naturally into discrete structural domains. The principal clue was that the fault broke along nearly the entire length of the active mountain front. Only along the ~90-km length of the 1999 rupture does the flank of the Western Foothills sport such a sharp, youthful, pre-existing thrust scarp [*Chen et al.*, 2002] (Figure 10). The length of this young scarp also coincides with the lengths of the Taichung piggyback basin and the tablelands.

The northern termination of the 1999 rupture occurred in a region where the structural and geomorphic expression of the thrust belt changes markedly [*Chen et al.*, 2000; *Lee et al.*, 2000]. North of the 1999 rupture, the tablelands dramatically change character and the Taichung foreland basin ends. All these observations indicate that there are fundamental structural differences between the portion of the Western Foothills and foreland basin adjacent to the rupture of 1999 and the regions to the north.

Much has been written about the Chelungpu fault, especially since its rupture in 1999. This structure may, in fact, have become the most thoroughly studied thrust fault on Earth, given the treasure trove of seismologic, geodetic, geomorphic and structural data that was mined in the aftermath of the earthquake. Here we summarize only that subset of information relevant to defining the Taichung Domain.

It is clear from structural analyses that the Chelungpu is the principal late-Quaternary thrust fault of the Taichung Domain. All agree that the fault dips shallowly east under the Western Foothills [e.g., *Lee*, 1949; *Meng*, 1963; *Pan*, 1967;

Chang, 1971; *Namson*, 1981]. The fact that the fault plane is at the base of and parallel to bedding of the Pliocene Chinshui Shale indicates that the fault dip must shallow at a depth of about 6 km into a bedding-plane detachment [*Namson*, 1981; *Y.-G. Chen et al.*, 2001a]. Total dip-slip offset on the fault must be at least 5000 m at the latitude of Taichung Basin to have produced this relationship. Hundreds to thousands of nominal Chi-Chi ruptures would have been necessary to accumulate this offset.

Pre-existing scarps along the 1999 rupture are strong evidence for the repetition of rupture of the Chelungpu fault in the Holocene and late Pleistocene [*Chen et al.*, 2000; *Chen et al.*, 2002]. Paleoseismological research indicates that the penultimate event occurred between 200-800 years ago [e.g., *W.-S. Chen et al.*, 2001a, b, c; *Lee et al.*, 2001b; *Ota et al.*, 2001a, b], and at least at one site vertical offset was nearly identical to the ~1.5 meter experienced in 1999 [*Streig et al.*, 2005].

Geodetic measurements made in the years prior to the earthquake [e.g., *Yu et al.*, 1997; 2001] yielded strain rates of about 10 mm/yr across the Taichung Domain, far lower than contemporaneous rates across the Chiayi Domain to the south. This relatively low pre-seismic geodetic rate led some to the conclusion that the Chelungpu fault was inactive [e.g., *Hu et al.*, 2001b]. However, even a horizontal strain accumulation rate of 10 mm/yr would suggest repetitions of 1999-like events every few hundred years [e.g., *Dominguez et al.*, 2003].

Deformed river terraces in the hanging-wall block of the Chelungpu fault reveal the presence of active fold scarps above kink bands associated with changes in dip of the fault. The most prominent of these is a fold scarp that traverses terraces of the Tachia River, about 6 km east of the northernmost section of the Chelungpu fault (Figure 10). Another fold scarp clearly disrupts terraces of the Wu River about 3 km east of the fault. These scarps were known before the 1999 event, but were thought to be secondary fault scarps [e.g., *Ku*, 1963; *Shih et al.*, 1985b, 1986; *Yang*, 1986]. Detailed investigation after the earthquake revealed that they are monoclinal flexures produced by movement of the hanging-wall block through a change in the dip of the underlying Chelungpu fault [*Suppe et al.*, 2001; *Lai*, 2002].

The other major active fault of the Taichung Domain underlies the Taichung Basin and the tablelands to the west (Figure 10). This fault bears several names in the literature [e.g., *Shih et al.*, 1983b, 1984b, c; *Shih and Yang*, 1985; *Yang*, 1986], but is commonly referred to as the Changhua fault.

The tablelands are the geomorphic expression of the Changhua fault (Figure 10). Their surfaces and underlying Quaternary deposits form an anticlinal welt separating the Taichung and coastal foreland basins. Along the southern portion of the Pakua Tableland, its western limb is much steeper and shorter than its eastern limb [*Deffontaines et al.*, 1997; *Ota et al.*, 2002]. To the north, the anticline is more symmetrical, with its axis running along the center of the tableland. Small steep west-facing scarps along the western flank of the Houli and Tatu Tablelands suggest that the Changhua fault breaks the surface, at least locally [*Pan et al.*, 1983; *Shih et al.*, 1983b, 1984b, c; *Shih and Yang*, 1985; *Yang*, 1986]. Additionally, sharp, minor east-facing scarps east of, and parallel to, the major scarp suggest that minor back-thrusts or kink bands are present at least locally in the hanging-wall block of the Changhua fault [*Lin*, 1957; *Ku*, 1963; *Sun*, 1965; *Tang*, 1969; *Shih et al.*, 1983b, 1984b, c; *Shih and Yang*, 1985; *Yang*, 1986]. Despite these steep scarps, several recent seismic reflection lines suggest that the Changhua fault itself may be a fault-propagation fold that does not break cleanly to the surface [*J.-C. Lee et al.*, 1997a; *Shih et al.*, 2000, 2002; *Yang et al.*, 2000; *Hsu*, 2001], as was suggested by *Chen* [1978]. Nonetheless, the presence of steep scarps suggests that deformation on the western flank of the tablelands could be severe locally during ruptures of the Changhua fault.

On the southern part of the Pakua Tableland are several E-W striking scarps (Figure 10). Although some have suggested that these scarps represent active faults [e.g., *Tomita*, 1932; *Mouthereau et al.*, 1999; *Tsai and Sung*, 2000], others consider them to be fluvial terrace risers [*Ku*, 1963; *Shih and Yang*, 1985; *Yang*, 1986]. The latter interpretation is more plausible, because the scarps are approximately parallel to the flow direction of the nearby Choushui River, and the lateritic soils on top of surfaces separated by those scarps are less well developed on the lower surfaces [e.g., *Chien*, 1993; *Shyu*, 1994].

The Changhua fault appears to represent the western limit of active deformation at the latitude of the Taichung Domain. Thus, the western limit of deformation in the Taichung Domain is substantially farther to the east than it is in the neighboring Chiayi and Miaoli Domains (Figure 3). Its surface trace lies approximately above a major, inactive normal fault [e.g., *Pan*, 1967; *Hsiao*, 1968; *Chang*, 1971; *Chou*, 1971; *Chen*, 1978]. It is not clear that the two faults are coincident at depth, but if they are, the Changhua fault would be a thrust fault that has re-activated a Tertiary normal fault, a phenomenon that appears to widespread in the western Taiwan fold and thrust belt [*Suppe*, 1984; 1986].

Total reverse slip on the Changhua fault is merely a few hundred meters [*Chen*, 1978], far less than the many kilometers of slip on the Chelungpu fault. However, the long-term slip rate and the age of initiation of the Changhua fault are unknown currently, so that the fraction of strain across the Taichung Domain that is taken up by the Changhua fault is not known. Nor have paleoseismic studies shed any light on the recurrence characteristics of the fault.

Both ends of the 1999 rupture provide important information about the transitions of the Taichung Domain with its neighbors to the north and south. In the south, geomorphic features indicate that both the Chelungpu fault and the Changhua fault diminish in prominence south of the Choushui River and terminate in the dextral strike-slip fault mentioned earlier. Along the Chelungpu fault, the 1999 rupture became more obscure south of the Choushui River [*Central Geologic Survey*, 1999a, b] (Figure 10). The geomorphic expression of the Chelungpu and Changhua faults also diminishes in this region. Some maps correlate the monoclinal western margin of the Pakua Tableland with the western flank of the Touliu and Chiayi Hills to the south [e.g., *Chinese Petroleum Corporation*, 1982, 1986]. However, the fundamentally different geomorphic expression of these two hills and structural form of the underlying sediments suggests that this correlation is inappropriate. Instead, the western flank of the Pakua Tableland, and by association the underlying tip of the Changhua fault, probably extend along scarps on the east side of the Chinshui River valley (Figure 10). Therefore, the Changhua fault probably diminishes southward along this tributary river valley and merges with the Chelungpu fault as shown in Figure 10 [*Ota et al.*, 2002].

This interpretation suggests an overlap in structures of the Taichung and Chiayi Domains. Structures underlying the Touliu and Chiayi hills are the northernmost expressions of the Chiayi Domain, and the dying southern strands of the Chelungpu and Changhua faults south of the Choushui River are the southernmost expressions of the Taichung Domain. The overlap of these structures indicates that the transition between the two domains is about 20 kilometers wide (Figure 3).

Complexities at the northern end of the Taichung Domain define a wide transition zone with its northern neighbor, the Miaoli Domain. We have already noted that the trace of the 1999 rupture ends near the northern end of clear topographic definition of the Chelungpu fault. In addition, the dip of the fault shallows northward, in concert with the shallowing dip of the Chinshui Shale, and the active fault dies out into a broad syncline to the north [*Lee et al.*, 2000; *J.-C. Lee et al.*, 2002]. The odd, east-striking anticlinal welt and bounding thrust faults and kink bands that formed the ~10-km-long northern segment of the 1999 rupture [*Y.-G. Chen et al.*, 2001a] form a small rabbit-ear-like anticline that cuts across regional bedding and into this syncline. Spectacular though this co-seismic welt was, it is not a large regional structure. Although *Ji et al.* [2001] and *Johnson et al.* [2001] considered this feature to extend to a depth more than 10 km in their geodetic models, the surficial deformation along this sector of the rupture suggested that these folds and faults are minor and shallow [*Lee et al.*, 2000; *J.-C. Lee et al.*, 2002].

The Chihhu syncline, across which this minor appendage of the 1999 rupture cut, is a major element of the Miaoli Domain to the north (Figures 10 and 11). This indicates that the northernmost rupture of the Chi-Chi earthquake is not really part of the Chelungpu fault; rather it is a surficial accommodation structure within the transition zone between the Taichung and Miaoli Domains.

Another indication of the broad transition between the Taichung and Miaoli Domains are the complexities of the Sanyi fault (Figure 10). Structurally, this fault is the principal northern extension of the Chelungpu fault [*Meng*, 1963; *Chinese Petroleum Corporation*, 1982, 1994; *Chen et al.*, 2000]. It ends in a sharp curve near the town of Sanyi that is perhaps analogous to the northern curve of the 1999 rupture. Minor geomorphic features indicate that this fault system has experienced minor slip in the recent past [*Tang*, 1969; *Yang*, 1986; *Lee*, 1994].

A co-seismic fault rupture, associated with a large earthquake in 1935, is also within the transition zone between these two domains. The predominantly right-lateral Tuntzuchiao fault extends southwestward across the northern edge of the Taichung Basin into the Houli and Tatu Tablelands. Vertical and right-lateral displacements were 0.5 to 1 meter and 1.5 to 2 meters, respectively [*Otuka*, 1936; *Bonilla*, 1975, 1977; *Chang*, 1976] (Figure 10). In fact, the fault appears to separate distinct deformational styles of these two tablelands. The topographic expression of this fault has been mostly obliterated, but the location of the 1935 rupture is known from the detailed map of *Otuka* [1936], and some small landforms are still visible on the 40-m DEM. Furthermore, the location of the fault is also constrained by several geophysical studies [e.g., *Pan et al.*, 1983; *Wang et al.*, 1994]. The Tuntzuchiao fault cuts across the strike of the basin and tablelands at about a 45˚ angle, about the relationship one would expect for a strike-slip structure serving as an accommodation structure within a fold-and-thrust belt [e.g., *Yeats et al.*, 1997]. The fault separates the tablelands to the south from the broad open Tunghsiao anticline and associated folds of the Miaoli Domain to the north (Figures 10 and 11). The width of the transition zone between the Taichung and the Miaoli Domains, as measured by the distance between the southern end of the Tuntzuchiao fault and the northern end of the Sanyi fault, is about 20 kilometers.

Miaoli Domain

The active structures of the Miaoli Domain are roughly on trend but contrast sharply with those of the Taichung Domain (Figure 3). The active tectonic landforms on land are broad coastal and narrower foothill anticline-syncline pairs (Figure 11). Reverse faults, including one that ruptured during an earthquake in 1935, are locally clear in the topography. The fact that active folding extends to the west coast suggests that a west-verging master thrust fault underlies the entire region between the Western Foothills and the coast. GPS geodetic measurements show that convergence across the domain is approximately N40˚W at about 5 mm/yr [e.g., *Yu et al.*, 1997], similar in azimuth but far lower in magnitude than convergence across domains to the south.

The narrow Miaoli valley occupies a position analogous to the Taichung Basin, farther south, in that it separates the two major geologic units of the domain (Figure 11). To the east, Mio-Pliocene, shallow continental-shelf sediments compose the Western Foothills [*Chou*, 1976]. The lower hilly areas to the west are mostly underlain by Quaternary post-orogenic sandstones and conglomerates, and are capped by deformed fluvial surfaces cut by small, recent fluvial valleys.

The two major anticline-syncline pairs that dominate the landforms of the Miaoli Domain are along strike with the features of the Taichung Domain. The eastern pair, the Chuhuangkeng anticline and Chihhu syncline, forms the western flank of the Western Foothills. The anticline is a very prominent landform (Figure 11). Its steep limbs strike about N20˚E and demarcate one of the major oil and gas fields in Taiwan. The Chihhu syncline parallels the anticline but broadens to the south. Its axis appears to be contiguous with an active east-verging kink band that deforms fluvial terraces of the Tachia River farther south, on the hanging-wall block of the Chelungpu fault, as discussed above (Figures 10 and 11).

The Shihtan fault, a bedding-plane back-thrust on the steep eastern limb of the syncline, ruptured during the 1935 earthquake [*Hayasaka*, 1935; *Otuka*, 1936; *Wang et al.*, 1994]. This fault dips at about 70 to 80 degrees to the west, and sustained vertical offset of about 3 meters during the earthquake [*Hsu and Chang*, 1979; *Pan et al.*, 1983]. Geomorphic and stratigraphic evidence suggests that the fault had been active before the earthquake [*Bonilla*, 1975; *Pan et al.*, 1983]. The structural position of the Shihtan fault, on the limb of a syncline and parallel to bedding, indicates that it is not a major structure.

West of the Miaoli valley is the other pair of major folds, the Tunghsiao anticline and Tunglo syncline [e.g., *Chang*, 1974; *Chinese Petroleum Corporation*, 1994]. In contrast to the Chuhuangkeng anticline, the Tunghsiao anticline is broad, and its flanks dip gently. The geomorphic expression of this anticline is less clear than that of the Chuhuangkeng anticline, because only gently dipping, less consolidated young beds are exposed in its eroded core. Tilted remnants of a lateritic terrace west of the town of Sanyi form a prominent cap to the eastern limb of this anticline [e.g., *J.-C. Chang et al.*, 1998]. The Tunglo syncline traverses the valley that runs northward from Sanyi [*Chinese Petroleum Corporation*, 1994]. The steepness of the hillside west of Sanyi suggests that a minor reverse fault separates the anticlinal and synclinal limbs there [*Y. Ota*, unpublished data].

In contrast to the Taichung Domain, no major thrust fault crops out on land within the Miaoli Domain. Even the largest historical event in this domain, the destructive M 7.1 earthquake of 1935, involved rupture of relatively minor faults: the Shihtan fault, on the limb of the Chihhu syncline, and the Tuntzuchiao fault, a right-lateral accommodation structure within the transition zone between the Miaoli and Taichung Domains (Figure 11). Nonetheless, the two major anticline and syncline pairs probably are the superficial manifestation of a major underlying detachment [e.g., *Elishewitz*, 1963; *Suppe and Namson*, 1979; *Namson*, 1981, 1983, 1984; *Hung and Wiltschko*, 1993]. This major detachment, if it were to crop out at the surface, would do so offshore and delimit the western edge of the Miaoli Domain. In fact, previous seismic reflection surveys have found folds offshore of the Miaoli area [*Wang*, 1967], indicating that the detachment does indeed continue beneath the offshore region.

The fact that the deformation front of the Miaoli Domain is much farther west than that of the Taichung Domain suggests a structural down-step from the Changhua and Chelungpu faults of the Taichung Domain to the décollement of the Miaoli Domain [e.g., *Y.-G. Chen et al.*, 2001a]. This step is accommodated by many minor structures in the transition zone between the two domains.

A dense NW-trending belt of seismicity (the Sanyi-Puli seismic belt) crosses the southern Miaoli Domain [*Lee*, 1994; *J.-F. Lee et al.*, 1997]. Based on a hypothesis that left-lateral structures are important active structural systems throughout western Taiwan [e.g., *Deffontaines et al.*, 1997; *Mouthereau et al.*, 1999, 2001], this seismic belt has been considered by some to be a major left-lateral fault, extending from the west coast to the Central Range (the Puli fault) [*C.-I. Lee et al.*, 1998]. However, no geomorphic evidence supports the existence of such a structure. Terrace surfaces above the seismic lineament do not exhibit sinistral disruption. Thus, we do not believe the Puli fault exists. Alternatively, the wide seismic belt may well indicate a zone of abundant minor fractures in the transition zone between the Taichung and Miaoli Domains. A more detailed analysis of the locations and mechanisms of this seismicity is needed to interpret its structural significance.

A series of E-W striking structures appear on some geologic maps between the Miaoli Domain and the Hsinchu Domain to the north [e.g., *Yang et al.*, 1996; *Chinese*

Petroleum Corporation, 1994]. Geomorphically, only the Touhuanping fault is well-expressed (Figure 11). This fault truncates the major folds both to the south and to the north, and is likely to be dominated by right lateral movement [e.g., *Tang*, 1968; *J.-S. Chen*, 1970, 1974; *Chiang*, 1970; *Tang and Hsu*, 1970; *Chang*, 1972; *Huang*, 1984; *Shih et al.*, 1985c; *Yang*, 1986; *Yang et al.*, 1997; *Shih et al.*, 1999]. Such E-W striking dextral faults have been suggested to be high-angle faults, possibly occupying pre-existing normal fault traces [e.g., *Chen*, 1972; *Lee et al.*, 1993]. These dextral structures are likely to also be accommodation structures between the detachment faults of the Miaoli and Hsinchu Domains. In places, deformation of young fluvial terraces indicates that these faults have been active recently [e.g., *Shih et al.*, 1985c; *J.-C. Chang et al.*, 1998].

Less than ten GPS stations are within or near the Miaoli Domain, and these show widely varying motions relative to the Penghu Islands [e.g., *Yu et al.*, 1997]. These stations appear to indicate N40˚W contraction across the region at about 5 mm/yr. This rate of strain accumulation is far lower than the rate in the Chiayi Domain, farther south. At this rate, the recurrence interval of a 5-m event would be about 1000 years, far longer than the proposed recurrence interval of the Chelungpu fault. This indicates that the crustal movement from the collisional processes is decreasing north of the Taichung Domain.

Hsinchu Domain

The Hsinchu Domain contains the northernmost evidence for active crustal shortening in Taiwan. It is a transitional domain between the active shortening of the Miaoli Domain and active extension of the Taipei Domain (Figure 3). Its reverse and strike-slip faults bring the active deformation front of the Miaoli Domain back onto land. N-S shortening across the domain is accommodated by two groups of reverse faults and folds. GPS measurements are not coherent across the Hsinchu Domain but do suggest that shortening rates are no more than a few mm/yr.

Most of the Hsinchu Domain is underlain by post- and syn-orogenic fluvial and marine sandstones and conglomerates. These were deposited in the foreland basin that formed atop the shallow continental shelf and west of the orogeny that produced the Western Foothills between the middle and late Pleistocene. Although numerous faults associated with that initial phase of the orogeny have been mapped within the shallow-marine continental shelf rocks of the Western Foothills [e.g., *Chinese Petroleum Corporation*, 1978], none of these appear to have produced geomorphic or stratigraphic evidence of activity in the past few hundred ky. Instead, the expressions of currently active structures lie farther west, in low hills that have risen recently from the coastal plain.

Two groups of active structures disrupt rocks of the foreland basin. Deformed geomorphic surfaces atop these sediments aid in defining their nature. Surfaces south of the west-flowing Touchien River are distinctly different from those to the north (Figure 12). North of the river is the Taoyuan-Hukou Tableland, a 25-km-square, largely uneroded parcel of land that consists predominantly of abandoned lateritic terraces of the Tahan River. The fact that older and higher terraces are in the south and younger and lower terraces are in the north and east demonstrates the northward migration of that river during the late Pleistocene. The youngest and northernmost of the Tahan River terraces formed about 32 ka [*Y.-G. Chen, 1988*]. The paleo-Tahan River was captured soon thereafter and now flows into the Taipei Basin [*Lin*, 1957; *Y.-G. Chen*, 1988; *Huang*, 1995] (Figure 13).

The northern of the two groups of active structures disrupts the surfaces of the Taoyuan-Hukou Tableland as a series of ENE-WSW striking scarps and fold axes (Figure 12). The scarps have long been considered to be active fault or fold scarps [e.g., *Hanai*, 1930; *Ku*, 1963; *Sunlin*, 1982; *Yang*, 1986]. The most prominent scarps are as much as

110 meters high and face each other across a 2-km-wide valley. All of the scarps are related to two major structures, the Pingchen and Hukou anticlines [e.g., *Tang*, 1963; *Wang*, 1964]. The Hukou anticline is likely to be a fault-bend fold created by an underlying thrust ramp, and the higher scarp south of the valley is a manifestation of its forelimb. The Pingchen anticline appears to be the result of a wedge thrust, and the lower scarp north of the valley may be a fold scarp related to its forelimb [*Wang*, 2003].

The two principal scarps become lower and less prominent to the east, where they traverse younger terrace surfaces. Earlier work suggested that the scarps terminate at the terrace riser between the youngest (Taoyuan) surface and the second youngest (Chungli) lateritic terrace [*Chan*, 1995]. However, a recent survey indicates that these scarps do extend onto the Taoyuan surface, only with a much smaller offset [*Wang*, 2003]. This new discovery demonstrates that the structures have been slightly active since formation of the Taoyuan surface, about 32 kyr ago.

A right-lateral fault zone that strikes WNW-ESE forms the southern boundary of the structures of the Taoyuan-Hukou Tableland [*Wang*, 2003]. Squeeze-ups and other geomorphic features along the fault zone clearly indicate a right-lateral sense of slip. In places this structure also appears to offset the above-mentioned scarps on the tableland. The fact that this right-lateral fault zone heads offshore suggests that it is a tear fault between the structures of the Tableland and unidentified structures offshore.

The other group of active structures in the Hsinchu Domain is geomorphically apparent just south of the Touchien River (Figure 12). Scarps of the Hsincheng and Hsinchu faults and enigmatic intervening scarps cut and deform the narrow fluvial terraces south of the river. The overall trend of the fault system is east-west.

The Hsincheng fault is expressed as a N50E-striking scarp that ranges in height from 50 meters on the oldest fluvial terrace to just a few meters on younger ones—a clear indication of the continuing activity of the fault during the period of formation of the river terraces. The existence of this scarp has long been known [e.g., *Tan*, 1934; *Ku*, 1963; *Tang*, 1968; *Shih et al.*, 1985c; *Yang*, 1986; *Chang et al.*, 1999]. On the bank of the Touchien River, an outcrop shows that this fault cuts the youngest terrace surface [*Lu et al.*, 2000; *Chen et al.*, 2004].

Subsurface evidence for the Hsincheng fault demonstrates that it is one of several major thrust faults in the western fold and thrust belt of Taiwan [e.g., *Pan*, 1965; *Tang*, 1968; *Tang and Hsu*, 1970; *Chang*, 1972; *J.-S. Chen*, 1974; *Hsu and Chang*, 1979; *Huang*, 1984; *Lu et al.*, 2000]. Moreover, *Namson* [1984] synthesized a variety of these stratigraphic data in a structural analysis that proposed this fault to be a major boundary thrust, with the décollement located within the Chinshui Shale, the same unit as is occupied by the major décollement of the Chelungpu fault [e.g., *Chang*, 1971]. It may be that only the short length of the Hsincheng fault that traverses the southern terraces of the Touchien River is currently active, however [*Chen et al.*, 2004]. The trace of the fault within the eroded hills southwest of the terraces displays no clear evidence of recent activity. If this reach is active, the rate of erosion would have to be equal to or greater than the rate of vertical motion across the fault.

Published evidence for the existence of the Hsinchu fault is less compelling than for the Hsincheng fault, so that this fault has not been considered to be an important active fault [e.g., *H.-C. Chang et al.*, 1998; *Lin et al.*, 2000a]. Most prior analyses consider it to be a minor high-angle reverse fault, reactivated along a pre-existing normal fault [*Pan*, 1965; *Tang*, 1968; *Chiu*, 1970; *Tang and Hsu*, 1970; *Chang*, 1972; *Shih et al.*, 1985c; *Yang*, 1986; *Lee et al.*, 1993; *Yang et al.*, 1994; *Lu et al.*, 2000]. The obvious fault scarp south of Hsinchu City, however, indicates that this fault is a major active structure. The fact that scarps related to the Hsinchu fault are at an angle with the coastline and far from the present-day riverbed indicates that they are unlikely to have been cut by the ocean or the river. Moreover, the Touchien River terraces immediately to the east of the scarps are anticlinally deformed and tilt upstream [*Chen et al.*, 2004]. Locally, small scarps are also present on very young river terrace surfaces [*Shih et al.*, 1985c; *Yang*, 1986].

The Hsinchu and Hsincheng faults appear to be linked by a 5-km-long zone of strike-slip faults that cut the terraces of the Touchien River [*Chen et al.*, 2004]. It is not surprising that previous workers [e.g., *Shih et al.*, 1985c; *Yang*, 1986; *Chang et al.*, 1999] mapped these scarps as fluvial terrace risers, because they are parallel to the Touchien River valley. However, two characteristics indicate that they may not be simple erosional scarps. First, they are not uniform in height and, in fact, are not even continuous. Second, anticlinal welts are associated with the scarps locally. Although there is no clear direct evidence for the sense of slip on these faults, their orientation relative to the Hsinchu and Hsincheng faults suggests that they are right-lateral, with a component of vertical motion, south side up.

Both of the two groups of faults in the Hsinchu Domain accommodate shortening in a direction that is nearly north-south. Thrust faults that strike NNE dominate the domain, but right-lateral faults intersect these at angles of about 45˚ [e.g., *Yang et al.*, 1996, 1997]. The direction of shortening across these faults is NNW, nearly 80˚ clockwise from the direction of shortening across the Miaoli Domain.

This kinematic interpretation is consistent with the Hsinchu Domain being the northernmost region of shortening in Taiwan. The plane separating the influence of the Manila and the Ryukyu subduction systems lies immediately to the north. To the south, the Miaoli Domain continues to suffer WNW-ESE shortening as a result of the collision of the forearc sliver of the Central Range and the South China continental shelf. To the north, the Taipei Domain is experiencing post-collisional collapse in the region of back-arc extension above the Wadati-Benioff zone of the Ryukyu subduction system [*Teng*, 1996; *Teng et al.*, 2000] (Figure 3). The neutral plane separating these two regimes is immediately north of the Hsinchu Domain. In this context it is reasonable to conclude that the structures of the Hsinchu Domain are merely the northern end of the structures associated with the major detachment of the Miaoli Domain, as suggested by *Biq* [1992]. However, bathymetric data offshore Hsinchu and Miaoli are insufficient to test this hypothesis [e.g., *Pan and Hu*, 1972].

As with the GPS vectors in the Miaoli Domain, those in the Hsinchu Domain are not coherent but suggest rates of shortening less than about 10 mm/yr [*Yu et al.*, 1997]. A longer period of monitoring will be required to establish a coherent pattern of strain and to assess whether or not the decadal geodetic pattern of strain is consistent with the patterns that are apparent from the geomorphology, stratigraphy and structural geology.

Taipei Domain

Within the Taipei Domain, formerly active thrust faults are being buried by young fluvial, marine, and volcanic sediment and locally obliterated by the construction of a volcanic edifice. NW-SE extension across the domain is concentrated on one principal normal fault, across which the long-term rate of extension appears to be at least 1 mm/yr (Figure 13). Motion across this structure, the Shanchiao fault, has been producing the Taipei Basin half-graben for the past few hundred thousand years. Seismic slip on the fault may have produced a marine incursion into the Taipei Basin in 1694. Decadal rates of extension across the domain are poorly constrained by GPS measurements [*Yu et al.*, 1997], but the magnitude of extension could be consistent with the long-term rate of extension across the Shanchiao fault.

The Taipei Domain comprises four physiographic regions: the Taipei Basin, the Linkou Tableland, the Tatun volcanic edifice, and the Western Foothills (Figure 13). The Western Foothills consist of Miocene marine rocks deposited on the continental shelf and subsequently folded and faulted during NW-SE shortening. The Linkou Tableland is a dissected flat lateritic surface [*Chen and Teng*, 1990; *Liew et al.*, 1990, 1991]. This ancient surface is underlain by up to 300 meters of undeformed fluvial sands and gravels, the oldest of which is several hundred thousand years old [*Liew et al.*, 1990, 1991]. A thrust fault that runs along the eastern flank of the tableland does not penetrate these deposits, so one can infer that shortening across at least that thrust fault had ceased before their deposition [*Teng et al.*, 2001].

The Tatun volcanic edifice, in the northern part of the Taipei Domain, consists of a series of late Quaternary volcanoes [e.g., *Chen and Wu*, 1971; *Juang and Bellon*, 1984; *Juang and Chen*, 1989; *Wang*, 1989; *Tsao*, 1994]. The predominance of andesitic volcanic rocks suggests a relation to Ryukyu arc volcanism [e.g., *Chen*, 1975, 1990], although certain geochemical characteristics suggest a post-orogenic extensional origin to some [e.g., *K.-L. Wang et al.*, 1999; *Chung et al.*, 2001]. The oldest andesitic rocks of the group are about 2.6 Myr old [e.g., *Wang*, 1989; *Chen*, 1990], but the bulk of volcanic production appears to have occurred about 200-300 Kyr ago [e.g., *Tsao*, 1994; *Song et al.*, 2000a, b]. Many of the craters and flows of the Tatun volcanic edifice remain well expressed geomorphically (Figure 13).

At the center of the Taipei Domain lies the Taipei Basin, home to the capital city of Taiwan (Figure 13). The basin is a half-graben, filled with sediments that lie upon Tertiary sedimentary rocks of the northern fold and thrust belt [*Teng et al.*, 2001]. Its fluvial fill has come from the major drainages that converge on the basin from the southwest, south, and northeast. The sedimentary fill is up to 679 meters thick near its northwestern edge [*Central Geologic Survey*, unpublished data] and thins to the south [*Peng*, 1998; *Teng et al.*, 2001]. According to luminescence dating of the deepest sediments of the basin, the filling of the basin with fluvial, estuarine and lacustrine sediments began about 400,000 years ago [*Peng*, 1998; *Wei et al.*, 1998].

The major active structure of the Taipei Domain is the Shanchiao fault, which separates the Taipei Basin and the Linkou Tableland (Figure 13). The observation of numerous triangular facets along the eastern foothills of the Linkou Tableland led to the discovery of the fault many decades ago [*Tan*, 1939; *Lin*, 1957]. The difference of more than 500 m between the elevations of basement rocks across the fault also points to its existence [e.g., *Wang-Lee et al.*, 1978; *Wang et al.*, 1995]. The fault also appears in shallow seismic reflection and refraction profiles [*Hsieh et al.*, 1991, 1992; *Sun*, 1990; *Hsiao*, 1993]. Recently, a dense array of drill cores has located the trace of the Shanchiao fault in the shallow subsurface more precisely [e.g., *Teng et al.*, 2001]. Also, *Yeh et al.* [1989a] and *Chen et al.* [1995] have interpreted small earthquake hypocenters to lie along the fault plane.

In general, the surface trace of the Shanchiao fault lies slightly east of the inactive Hsinchuang fault [*Sun*, 1990], a thrust fault that probably represents the deformation front of northern Taiwan in the early Quaternary Period [*Ho*, 1988; *Wang-Lee et al.*, 1978; *Teng et al.*, 2001]. The parallel traces of the two faults suggest that the steeper normal fault and shallower thrust fault merge at depth and that the younger Shanchiao fault has reoccupied the older fault plane [e.g., *Wu*, 1965; *Hsieh et al.*, 1992].

The Tatun volcanic edifice is clearly bisected by the northern extension of the Shanchiao fault (Figure 13). On the uplifted block, northwest of the fault, the semi-circular perimeter of about half the edifice is preserved. Southeast of the fault, the narrower width of the volcanic construct may well be due to burial of the flanks by sediments of the Taipei Basin. In detail, the scarp across the volcano is not as clear as it is between the Taipei Basin and Linkou Tableland. Nonetheless the overall trajectory of the fault zone over the volcano is clear, and numerous small scarps and other linear features form a clear fault zone. A recent fractal geometric analysis of the topography of the Tatun volcanic edifice revealed that the erosional rate northwest of the fault is higher than it is on the southeast, indicating that the northwest block of the fault is uplifting [*Yeh*, 1994]. The trace of the fault is also the locus of a linear distribution of hot springs and active fumaroles [*Chen and Wu*, 1971; *Tsao*, 1994; *Chu et al.*, 1998].

Farther northeast, west of Chinshan, the trace of the Shanchiao fault appears to follow a northeast-trending scarp in the volcano (Figure 13). A seismic reflection survey confirms the presence of the fault along this scarp [*Hsieh et al.*, 1991]. A scarp on the seafloor indicates that the fault continues offshore [*Yu and Shyu*, 1994; *Ma*, 1995].

History records a very interesting occurrence that is quite relevant to understanding the relevance of the Shanchiao fault to seismic hazard assessment. In 1694, historical documents indicate an abrupt subsidence in the Taipei region, with the formation of a brackish-water lake that inundated at least a third of the basin [*Shieh*, 2000]. The lake was deep enough to permit large ships to sail far into the basin. The formation of the lake is most easily explained by several meters of slip on the Shanchiao fault, since such a rupture would result in subsidence of a large portion of the Taipei Basin.

We have found no geomorphic evidence for other active structures within the Taipei Domain. Many bedrock faults crop out in the Western Foothills along the southern flank of the basin. Some of these have been considered active by others [e.g., *Hsu*, 1974; *Huang et al.*, 1994b; *H.-C. Chang et al.*, 1998]. However, the lack of geomorphic evidence of recent activity suggests that activity has ceased. The most recent version of the Central Geological Survey's active fault map, which does not include these faults, reflects this opinion [*Lin et al.*, 2000a].

The steep and tall escarpment along the southwestern flank of the Linkou Tableland is more difficult to dismiss (Figure 13). This feature has been attributed to normal offset along a fault that has been called the Nankan fault [e.g., *Hanai*, 1930; *Ku*, 1963; *Ho*, 1969; *Sunlin*, 1982; *Yang*, 1986; *K.-J. Chen et al.*, 1988; *Chen et al.*, 1994]. Others have suggested that it is not a fault scarp, but rather an erosional scarp formed by the Tahan river as it swept northward across the Taoyuan surface about 30,000 years ago [e.g., *Tsai*, 1986; *Shih et al.*, 1999; *C.-Y. Wang et al.*, 1999]. Certainly, the extreme difference in age of the Taoyuan surface and the thick lateritic Linkou surface [e.g., *Yang*, 1986; *Lin*, 1991] demonstrates that, even if it is a fault scarp, the height of the scarp does not represent the amount of vertical offset. More information is needed to resolve this debate, but the orientation of the feature suggests to us that it is not likely to be an active fault scarp.

The reoccupation by the Shanchiao fault of the Hsinchuang fault in the late

Quaternary Period demonstrates that the Taipei Domain has changed recently from a regime of shortening to one of extension. This is consistent with the movement of the domain out of the influence of the Manila subduction system to that of the Ryukyu subduction system [e.g., *Suppe*, 1984, 1987; *Teng*, 1996; *Teng et al.*, 2000]. The strain field of northern Taiwan has changed from a contraction-dominated environment to an extension-dominated one [*Lee and Wang*, 1988; *Lee*, 1989; *Wu*, 1994], wherein the extension is associated with the opening of the Okinawa Trough, the back-arc basin of the Ryukyu system [*Suppe*, 1984, 1987]. In marine seismic reflection sections offshore to the northeast of Taiwan, many old reverse faults have become normal faults in their uppermost parts, indicating the re-activation of these structures [*Hsiao et al.*, 1998; *B.-H. Hsu*, 1999].

Ilan Domain

The Okinawa Trough is the back-arc basin of the Ryukyu subduction system (Figure 1). Its progressive westward propagation is splitting the country's mountainous backbone along the western suture, now occupied by the Lanyang Plain (Figure 14). This large, triangular basin reflects the current foundering of the western tip of the westward propagating Okinawa back-arc trough (Figure 14). On the northwestern flank of the Lanyang Plain is the Hsueshan Range, consisting of late Eocene to early Miocene strata deposited probably in a submarine half-graben of the Asian continental shelf [*Teng*, 1992]. On the southern flank of the plain is the northern tip of the Central Range, a complex of late Paleozoic to late Mesozoic greenschist-facies metamorphic rocks and Miocene slates. Two observations suggest that these flanking ranges are sinking as well. First, their crests are appreciably lower than range crests farther south. Second, their flanks are being buried by alluvium on the east coast and in the Taipei Basin.

If *Suppe*'s [1987] calculation is correct that the western edge of Ryukyu subduction is migrating southwestward at about 100 mm/yr, then one would anticipate that none of the extensional features nor the sinking of the onshore part of the Ilan Domain would be older than about a half million years. This is supported by stratigraphic evidence that post-collisional collapse of northeastern Taiwan has occurred in just the past million years [*Teng*, 1996; *Teng et al.*, 2000]. Paleo-strain analyses also reveal a change from contraction to extension in the stress field of northern Taiwan [*Lee and Wang*, 1988; *Lee*, 1989].

Leveling and triangulation measurements suggest that the direction of extension across the Lanyang Plain is toward the SE, similar to the opening direction of the southern Okinawa Trough. They also suggest a fan-like opening of the plain about a vertical rotational axis at the western corner of the plain [*Sheu*, 1987; *Kuo and Yu*, 1994; *Liu*, 1995]. Recent GPS results indicate that the extensional direction is SE (139˚), with a maximum principal strain rate about 1 μstrain/yr [*Yu et al.*, 1995; *Yu and Chen*, 1996; *Y.-J. Hsu*, 1999]. Focal mechanisms of shallow earthquakes near Ilan and Kueishan Island also support a SE direction of extension [*Shen*, 1996; *Kao and Jian*, 1999].

Another important indication of recent back-arc activity near the Lanyang Plain is Kueishan Island (Figure 14), a small volcanic island just offshore. Its most recent eruption occurred less than 7 ka ago [*Y.-G. Chen et al.*, 2001b].

Two systems of geomorphically expressed normal faults are apparent in the onland portion of the Ilan Domain. They bound the Lanyang Plain on the northwest and on the south. Normal faulting along the northwestern flank of the Lanyang Plain is complex. Instead of possessing the simple morphology of a simple normal fault, the Hsueshan Range front displays a complex family of triangular facets that step up slope from the base of the range [*C.-T. Lee et al.*, 1998] (Figure 14). The scarps are especially prominent near the town of Toucheng. This family of short, discontinuous faults, marked by triangular facets, reflects the existence of a zone of normal faults several km wide.

The southern edge of the Lanyang Plain also appears to be controlled by normal faulting. The eastern half of this southern boundary, near Suao, is a very straight mountain front that is most simply interpreted as a fault scarp (Figure 14). The western half, however, shows no clear sign of a fault scarp. Instead, the valley edge consists of bedrock ridges protruding toward the valley and large alluvial fans extending up into the range. If there is an active normal fault along this section, sedimentation rates must be higher than faulting rates. Sediment cores from the plain reveal that this may indeed be the case. A sharp contact between bedrock and young sediment exists very close to the mountain front. Drilling just north of the mountain front encountered the top of bedrock at a depth about 40 meters. At a nearby site, only several hundred meters farther north, drilling to a depth of 200 meters did not encounter bedrock [*Chen*, 2000]. Hence, a steep scarp exists in the shallow subsurface between these two sites.

The seaward extension of these two normal fault systems is not very clear, and the lack of high-resolution bathymetry near the coast prohibits resolution of the issue (Figure 14). However, a submarine scarp along the northeastern extension of the northern normal fault system suggests that the northern structure continues for at least another 15 km. The presence of a submarine canyon just offshore Suao, on the other hand, indicates that the southern structure may not extend very far seaward.

Geophysical exploration sheds some light on the structure beneath the Lanyang Plain. Seismic reflection data, for example, may help explain why the faults on the northwest side of the plain are better expressed than those on the south. These data show that the sedimentary fill beneath the plain is asymmetrical. It is thickest along an axis striking about N60˚E and passing through Ilan [*Chiang*, 1976]. Along this axis, sediments just offshore reach thicknesses greater than 1500 meters. The fact that the gradient in sediment thickness is greater northwest of this axis than southeast suggests that rates of downdropping are greater on the faults to the northwest than on those to the south.

The axis of maximum sediment thickness is also nearly coincident with the axis of geodetically measured subsidence of the plain between Ilan and Luotung [*Earthquake Research Institute*, 1934]. Along this axis, the maximum subsidence rate is locally as high as 20 mm/yr [*Sheu*, 1987; *Kuo and Yu*, 1994; *Liu*, 1995]. Moreover, this axis of geodetic submergence coincides with a hypothetical structural line, suggested by both clusters of small earthquakes [*Tsai et al.*, 1975; *Tsai*, 1976] and a linear magnetic feature [*Yu and Tsai*, 1979]. However, whether this hypothetical line is an actual buried geologic structure remains unresolved.

The normal faults that bound the Lanyang Plain are generally considered to be the southwestward extension of normal faults bounding the Okinawa Trough. Bathymetry of the Okinawa Trough indicates that the most active opening area in the recent 2 million years is its southern part [*Hsu et al.*, 1996; *Wang et al.*, 1998], adjacent to the Lanyang Plain. The active opening is supported by many young normal faults shown in seismic reflection studies [e.g., *Herman et al.*, 1978; *Letouzey and Kimura*, 1986; *Sibuet et al.*, 1987, 1998; *B.-H. Hsu*, 1999], abundant seismicity [e.g., *Yeh et al.*, 1989b; *Sibuet et al.*, 1998], and high heat flow [e.g., *Lu et al.*, 1981] in that region.

The Lishan fault, which runs along the Lanyang River valley, upstream from the Ilan Plain, is not likely to be an active structure, despite its linear course. This fault is a major bedrock structure in Taiwan [e.g. *Ho*, 1986, 1988; *Yang*, 1986] (Figure 3), and it has a complicated structural history. It may have moved as both a normal fault and reverse fault, and has been reactivated at least once [e.g., *Teng et al.*, 1991; *Teng*, 1992; *J.-C. Lee et al.*, 1997b]. Geomorphic evidence, however, indicates that the fault is not likely to be active. Upstream of the small village of Tuchang, neither tributary valleys nor ridgelines show any disturbance by the fault (Figure 14). Downstream of Tuchang, the only active tectonic features form a swath of minor sub-parallel ridge-cutting scarps about five kilometers northwest of the valley floor. These are probably geomorphic evidence of active normal faults. These minor active normal faults are probably related to the incipient opening of the Lanyang Plain along the ancient Lishan suture.

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