INTRODUCTION

Curves in the topographic grain of active orogenic belts typically parallel faults and lithologic contacts in part because of linkages between uplift rates and the structural development of the overriding plate. However, in the Taiwan collision zone, the topographic grain trends nonparallel to mapped faults and folds in the central portion of the belt along the southwest flank of the Hsüehshan Range. Here, the northern side of the Puli topographic embayment trends ~345°, forming a topographic break that lies at a high angle to the structural grain but is nearly parallel to an underlying continental margin fracture zone in the downgoing plate. We analyzed fault-slip and other structural data, extracted normalized steepness indices from streams within the Tachia, Peikang, and Mei River basins, and integrated the results with recently published precise leveling data in order to understand the spatial and temporal variation in uplift and the structures that accommodated that uplift. Stress inversion reveals a NW-SE-trending maximum compression direction for an early-stage fault population and an ENE-WSW-trending maximum compression direction for a late-stage fault population. River steepness indices delineate a NW-trending boundary in incision rate that coincides with an increase in rock uplift rates derived from independent geodetic measurements. This NW-trending boundary is consistent with the late-stage maximum compression direction, suggesting that uplift of the Hsüehshan Range relative to the Puli topographic embayment has been accommodated by the late-stage faults. Our results suggest that a continental margin promontory has slowed underplating beneath the Puli topographic embayment and that the topographic grain of active collision zones is closely linked with the architecture of the downgoing plate.

Tectonic implications of nonparallel topographic and structural curvature in the higher elevations of an active collision zone, Taiwan

David C. Mirakian¹, Jean M. Crespi¹, Timothy B. Byrne¹, Chung Huang¹, William B. Ouimet¹, and Jonathan C. Lewis²

¹CENTRE FOR INTEGRATIVE GEOSCIENCES, UNIVERSITY OF CONNECTICUT, 354 MANSFIELD ROAD U-1045, STORRS, CONNECTICUT 06269, USA
²DEPARTMENT OF GEOSCIENCE, INDIANA UNIVERSITY OF PENNSYLVANIA, WALSH HALL ROOM 111, 302 EAST WALK, INDIANA, PENNSYLVANIA 15705, USA

ABSTRACT

Curves in the topographic grain of active orogenic belts typically parallel faults and lithologic contacts in part because of linkages between uplift rates and the structural development of the overriding plate. However, in the Taiwan collision zone, the topographic grain trends nonparallel to mapped faults and folds in the central portion of the belt along the southwest flank of the Hsüehshan Range. Here, the northern side of the Puli topographic embayment trends ~345°, forming a topographic break that lies at a high angle to the structural grain but is nearly parallel to an underlying continental margin fracture zone in the downgoing plate. We analyzed fault-slip and other structural data, extracted normalized steepness indices from streams within the Tachia, Peikang, and Mei River basins, and integrated the results with recently published precise leveling data in order to understand the spatial and temporal variation in uplift and the structures that accommodated that uplift. Stress inversion reveals a NW-SE-trending maximum compression direction for an early-stage fault population and an ENE-WSW-trending maximum compression direction for a late-stage fault population. River steepness indices delineate a NW-trending boundary in incision rate that coincides with an increase in rock uplift rates derived from independent geodetic measurements. This NW-trending boundary is consistent with the late-stage maximum compression direction, suggesting that uplift of the Hsüehshan Range relative to the Puli topographic embayment has been accommodated by the late-stage faults. Our results suggest that a continental margin promontory has slowed underplating beneath the Puli topographic embayment and that the topographic grain of active collision zones is closely linked with the architecture of the downgoing plate.
Differences in channel profile steepness were interpreted using precise leveling data recently obtained by Ching et al. (2011) and K.H. Chen et al. (2011) along the Mei River. Together, the structural and geomorphological studies suggest that the NNW-trending topographic break is structurally controlled. The region of nonparallel topographic and structural curvature in the higher elevations of central Taiwan lies above a continental margin promontory identified by Byrne et al. (2011) in the downgoing Eurasian plate. These authors proposed that the northern boundary of the promontory coincides with the NW-trending Sanyi-Puli seismic zone and is a continental margin fracture zone. This fracture zone and the NNW-trending topographic break are nearly coincident in map view and also lie at a low angle to each other. We conclude this paper by discussing the way in which this lateral variation in crustal architecture of the downgoing plate in central Taiwan may have affected collisional processes and the evolution of topographic expression in the mountain belt.

TECTONIC FRAMEWORK

The island of Taiwan is located on the boundary between the Philippine Sea plate and the Eurasian plate and consists of an active collisional mountain belt formed as a result of the impingement of the Luzon magmatic arc on the Eurasian continental margin beginning in Pliocene–Pleistocene time (Ho, 1986; Teng, 1990). To the east of Taiwan, the Philippine Sea plate subducts along the north-dipping Ryukyu arc-trench complex, whereas to the south of Taiwan, the South China Sea underthrusts the Philippine Sea plate in an east-dipping subduction zone at the Manila Trench (Teng, 1990).

Currently, the convergence vector of the Philippine Sea plate relative to the Eurasian plate is ~306° azimuth at ~8.2 cm/yr as determined by global positioning system (GPS) observations (Yu et al., 1997). The oblique nature of convergence between the N-trending Luzon arc and the NE-trending Eurasian continental margin has resulted in the southward propagation of the collision (Suppe, 1981). The orientation of the Eurasian margin and kinematics of the orogen result in precollisional tectonics to the south of the Hengchun Peninsula in the offshore accretionary prism, active collision in the central and southern portions of the exposed mountain belt, and postcollisional tectonics in northern Taiwan (Teng, 1990).

In the active collision in the central and southern portions of Taiwan, Byrne et al. (2011) have proposed that a continental margin promontory in the downgoing Eurasian plate is delineated on the north by the Sanyi-Puli seismic zone and on the east by a NE-trending magnetic anomaly high (Fig. 2). The Sanyi-Puli seismic zone is approximately ~100 km long and contains two distinct seismogenic layers at ~5–13 km and 20–33 km depth (Wu and Rau, 1998; Chen and Chen, 2002). On the basis of the northwest trend of seismicity, current relative plate convergence, and a number of strike-slip focal mechanisms, the Sanyi-Puli seismic zone has been interpreted as a left-lateral transfer fault accompanied by transpressive thrusting (Defontaines et al., 1997; Wu and Rau, 1998). Geophysical surveys by Hsu et al. (1998) have revealed a linear magnetic anomaly high that extends from central Taiwan to the South China Sea basin. In the South China Sea basin, the magnetic anomaly high coincides with a change...
in lower-crustal velocity identified previously by Nissen et al. (1995) using deep-penetration seismic-refraction surveys. Specifically, a thick layer of high-velocity crust between ~15 km and 30 km depth where P-wave velocities increase to >7.0 km/s was interpreted by Nissen et al. (1995) to consist of magmatically underplated gabbros that cooled at the rifted edge of the Eurasian continental margin. The magnetic anomaly high, thus, serves as a proxy for the edge of Eurasian continental crust (Byrne et al., 2011).

In map view, the edges of the continental margin promontory are also roughly outlined by a convex high-seismicity boundary with a low-seismicity core (Lin, 2001; Wu et al., 2004).

Horizontal GPS vectors diverge as crust to the east is forced around the low-seismicity region (Chang et al., 2003), which suggests that the crust in the promontory is mechanically stronger than crust to the east.

The southern and central part of the continental margin promontory includes the Peikang basement high, which lies between two laterally offset rift basins (Byrne et al., 2011), the Taihsi and Tainan basins (Lin et al., 2003). The continental margin fracture zone, which defines the northern boundary of the promontory, is delineated both by the Sanyi-Puli seismic zone and by the truncation of the magnetic anomaly high along the Sanyi-Puli seismic zone just north of the Puli basin (Fig. 2A; Byrne et al., 2011). In this interpretation, the Puli topographic embayment lies above the inferred eastern corner of the continental margin promontory, and the Hsüehshan Range lies directly north of the northern boundary of the promontory.

The Hsüehshan Range and Puli topographic embayment are underlain by Eocene–Oligocene slates of the Hsüehshan belt (Fig. 1). These slates belong to the slate belt, a major lithotectonic unit in Taiwan, which also includes Miocene slates of the Lushan Formation in the western Backbone belt and Eocene slates of the Pilushan Formation in the eastern Backbone belt (Ho, 1986). The Lishan fault separates the Hsüehshan belt on the west from the Backbone belt on the east.

The major lithotectonic units that lie to the west of the slate belt are the fold-and-thrust belt (Fig. 1). The fold-and-thrust belt is composed of imbricate thrust sheets and folds that expose unmetamorphosed to zeolite-grade Miocene–Pleistocene rocks (Ho, 1986). The Chuchih fault, also known as the Shuilikeng fault in central Taiwan, separates the fold-and-thrust belt on the west from the Hsüehshan belt on the east (Ho, 1986).

The major lithotectonic units to the east of the slate belt are the pre-Tertiary metamorphic complex and the magmatic arc (Fig. 1). The pre-Tertiary or Tananao metamorphic complex represents the unroofed Eurasian basement and is chiefly composed of schists and marbles metamorphosed to greenschist- and locally amphibolites-facies conditions (Ho, 1986; C.H. Chen et al., 1983). The Lazon magmatic arc belongs to the Philippine Sea plate and is separated from rocks of Eurasian affinity in part by the Longitudinal Valley fault.

GEOLOGY OF THE HSÜEHSHAN RANGE AND PULI TOPOGRAPHIC EMBAYMENT

The Puli topographic embayment is a topographically depressed zone ~40 km east of the frontal thrust of the orogen (~24°N, 121°E), and it is accompanied by a string of five basins that collectively constitute the intramountain Puli basin chain. To the north and east of the Puli basin chain, the Hsüehshan Range rises to >3500 m, extending for ~150 km along the island’s length. The NNW-trending topographic break that separates the Puli topographic embayment from the southwestern flank of the Hsüehshan Range can be delineated by a change in hillside values from less than to greater than 10° averaged over a 1-km-resolution grid. The trend of the topographic break is ~345°, which is only slightly oblique to the continental margin.
fracture zone currently reactivated as the Sanyi-Puli seismic zone (Fig. 2), but which is highly oblique to mapped faults and folds.

Although the Puli topographic embayment and the Hsiuehshan Range are physiographically distinct, as noted previously, both regions are floored by Eocene–Oligocene strata of the Hsiuehshan belt. The rocks in the Hsiuehshan belt are composed of a thick package of Oligocene strata with minor Eocene and Miocene members (Chen, 1977; Teng et al., 1991). According to Teng et al. (1991), these sediments were deposited in the Hsiuehshan trough, a Paleogene depositional basin occupying a fault-bounded half graben that formed during the rifting of the South China Sea. Strata are primarily composed of interbedded pelites and quartzitic sandstones, a reflection of the rapid changes in sea level and subsidence within the Hsiuehshan trough.

**Stratigraphy**

In the Puli topographic embayment and to the north in the Hsiuehshan Range, the major Eocene–Miocene strata that compose the Hsiuehshan belt are the Tachien Sandstone, Chiayang Formation, Paileng Formation, and Shuichangliu Formation. The Tachien Sandstone overlies the Eocene Shihpachunghsi Formation and is composed of massive fine- to coarse-grained quartzitic sandstone interbedded with shale and slate (Chen, 1977). The Chiayang Formation conformably overlies the Tachien Sandstone and is characterized by a thick series of slate with minor sandstone and siltstone interbeds. Biostratigraphic zones are difficult to identify in both the Tachien Sandstone and Chiayang Formation because of the poor preservation of microfossils. However, bivalves have been observed at a number of localities within the Tachien Sandstone and are probably Eocene in age (Tan, 1944), and some fossils date the Chiayang Formation to the Oligocene (Chang, 1971). Gradually thinning to the west, the Chiayang Formation interfingers and is replaced by the Paileng Formation (Chen, 1979). The Paileng Formation is composed of thick, white to gray, quartzitic sandstone interbedded with argillite and slate and has been tentatively assigned an Oligocene age based on stratigraphic relations (Chen, 1977). In contrast to the more indurated rocks to the east, the late Oligocene to early Miocene Shuichangliu Formation is very weakly metamorphosed and is chiefly composed of black argillite and shale.

**Metamorphism**

The metamorphic history and degree of induration of rocks in the Hsiuehshan belt markedly vary along strike. Strata in the northernmost Hsiuehshan belt record a history of dynamic metamorphism associated with the Pliocene–Pleistocene collision (C.T. Chen et al., 2011), as indicated by illite crystallinity and peak metamorphic temperature, which increases from west to east (C.H. Chen et al., 1983; C.T. Chen et al., 2011). In contrast, peak metamorphic temperature and metamorphic grade in the central and southern portions of the Hsiuehshan belt increase down section within the older stratigraphic units and do not systematically increase from west to east. This observation has been interpreted to reflect a history of burial metamorphism (C.T. Chen et al., 2011), which is further supported by the zonation of metamorphic facies.

Illite crystallinity studies show that metamorphic isograds are regionally folded with stratigraphy in the southern portion of the physiographic Hsiuehshan Range (Fig. 3; C.H. Chen et al., 1983). This can be observed where the greenschist-facies isograd is folded along with the Tachien Sandstone because the first occurrence of biotite is down section completely within this formation (Yen, 1973). As a result, the greenschist-facies isograd can be used to trace the exposure of the Tachien Sandstone, which crops out at the core of two structural highs, the Pahsinlu and Tachien anticlines. On the north side of the Puli basin, the greenschist-facies isograd and Tachien Sandstone are exposed, whereas on the south side, submetamorphic to zeolite-grade rocks from the younger Paileng Formation are exposed. The along-strike change in the degree of metamorphism on either side of the fracture

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**Figure 3.** Metamorphic facies map of west-central Taiwan. Circles, squares, and triangles—sampling localities for illite crystallinity analyses. Figure is modified from C.H. Chen et al. (1983).
zone suggests that rocks in the Hsüehshan Range were exhumed from deeper structural levels than rocks that lie to the south above the continental margin promontory. C.H. Chen et al. (1983) separated this offset in metamorphic grade with a NW-trending boundary and proposed the boundary may be structurally related (Fig. 3).

Structure

The Hsüehshan Range is generally recognized as a pop-up structure bounded to the east by the steep, west-dipping Lishan fault and to the west by the east-dipping Shuilikeng fault (Clark et al., 1993; Lee et al., 1997; Lu et al., 1997). Based on structural-stratigraphic studies, Teng et al. (1991) concluded that the Lishan fault bounded the Hsüehshan trough on its eastern side as a normal-displacement fault and that during the Phiocene–Pleistocene collision, the Lishan fault was reactivated as a back thrust. The Shuilikeng fault is poorly resolved by seismic-reflection data and is interpreted to dip steeply toward the east, possibly extending into the middle crust (Wang et al., 2002; Camanni et al., 2011). Although published maps from the Central Geologic Survey of Taiwan indicate that the Shuilikeng fault is not active, river incision studies from the Wu and Peikang Rivers, which cut through the Shuilikeng fault zone, indicate that the fault may have been active during the Holocene (Sung et al., 2000; Yanites et al., 2010).

Folds near the core of the Hsüehshan Range are typically open to tight with subvertical to steeply SE-dipping axial planes; fold axes trend northeast and are roughly subhorizontal (Clark et al., 1993; Tillman and Byrne, 1995). Metamorphic folds are normally cylindrical and also slightly asymmetric, with somewhat steeper NW-dipping forelimbs. Bedding-parallel faults with downdip slickenlines and fibrous steps yielding a thrust sense are present on both limbs of the fault. Bedding-parallel faults associated with flexural-slip folding accompanied shortening (Clark et al., 1993; Tillman and Byrne, 1995; Fisher et al., 2002).

A penetrative axial-planar slaty cleavage and downdip stretching lineation are well developed in the finer-grained rocks that compose the Chiayang and Paileng Formations (Tillman and Byrne, 1995). Cleavage commonly cuts bedding-plane faults associated with flexural slip, implying that cleavage formation postdated folding, possibly when fold thickening became an inefficient mechanism to accommodate shortening (Tillman and Byrne, 1995). Pressure shadow fibers related to cleavage formation record dominantly pure shear, plane strain deformation in the Hsiuehsien belt, which contrasts with the noncoaxial strain history of the western Backbone belt (Tillman and Byrne, 1995). Tillman and Byrne (1995) proposed that the coaxial strain history can be explained by the Lishan fault functioning as a steeply dipping, rigid backstop. Finally, late-stage, conjugate, strike-slip faults cut cleavage and bedding-parallel faults and have an approximately NW-trending σ 1 (Tillman and Byrne, 1995).

Based on the available stratigraphic, metamorphic, and structural data that have been collected from field sites located close to the core of the Hsüehshan Range, the Hsüehshan Range records a history of burial metamorphism followed by folding, limb tightening, cleavage development, and conjugate strike-slip faulting (Fisher et al., 2002). In the following section, these data are compared to structural and kinematic data that we collected from field sites located along the NW-trending topographic break in the southwest Hsüehshan Range.

ANALYSIS OF STRUCTURAL FEATURES ALONG THE TOPOGRAPHIC BREAK

Field data were collected along the 40-km-long topographic break between 24°00'N and 24°20'N latitude and 12°05'W and 121°02'W longitude (Fig. 4). The data consist primarily of fault measurements but also include measurements of the orientation of bedding and cleavage. Four rivers and one road that cross the mountain front were used to gain access to outcrops and are referred to as structural stations. From north to south, these are the Daan River, Da Hsüehshan Forest Road, Tachia River, Peikang River, and Mei River stations. At each station, field sites were selected based on their proximity to the topographic break, accessibility, and the relative quality of the exposure to establish whether the outcrops were in place. In addition, field sites were distributed along the length of the topographic break in order to minimize spatial bias that could complicate structural analyses. All of the field sites were located in low-grade, Eocene–Oligocene sandstone and slate, with the exception of a few field sites located along the Daan River and Da Hsüehshan Forest Road, which were in submetamorphic Miocene sandstone and shale.

Bedding and Cleavage Data

At the majority of field sites, the interbedded sequences of shale/slate and sandstone allowed for the easy identification of bedding surfaces. In total, 154 bedding measurements were compiled for the field area: 109 was measurements made during field work for this study, and 45 were obtained from the geologic map of Lo et al. (1999). A cylindrical best fit of poles to all the bedding measurements yields a mean fold axis that plunges gently toward the southwest at ~26° (Fig. 5A). In addition, field measurements of fold axes and the bedding-cleavage intersection lineation show that they consistently plunge toward the SW along the 40 km length of the topographic break.

Cleavage morphology varies significantly within the field area depending on the degree of metamorphism and grain size of the strata. Miocene rocks at field sites to the west of the Shuilikeng fault near the Daan River and Da Hsüehshan Forest Road lack even locally developed cleavage zones. However, immediately to the east of the Shuilikeng fault, a continuous, axial-planar slaty cleavage (S 1) is ubiquitous within the fine-grained strata of the Chiayang Formation. A strong cleavage is also present within the slaty layers of the Tachen Sandstone, and pencil cleavage was observed along the Peikang River in the Paileng Formation. In total, 30 measurements of cleavage were recorded at field sites where cleavage is continuous and consistently oriented. On average, cleavage dips steeply toward the SE at 76° (Fig. 5B).

The uniform southwest plunge of folds along the topographic break sharply contrasts with fold geometries observed in the core of the Hsüehshan Range, where bedding data indicate that fold axes are roughly subhorizontal (Tillman and Byrne, 1995). Folds in the field area may have been originally horizontal and later tilted because of differential uplift between the central and southern portions of the Hsüehshan Range. Under this scenario, regional-scale tilting may have also reoriented older brittle structures, and this possibility must be accounted for when analyzing fault-slip data collected near the topographic break.

Fault-Slip Data

In total, 104 faults were identified in the field, 72 of which contained both reliable striations and slip-sense indicators. Slip sense was inferred using a variety of indicators, including quartz slickenfibers, Riedel shears, tool marks, drag folds, and offset features such as tension cracks, bedding, and ripple marks (Fig. 6). The quality of slip-sense indicators was assessed in outcrop and confidently determined for all of the 72 faults with full geometric criteria. Of the 72 faults, 49 are thrust faults, 17 are strike-slip faults, and 6 are normal faults (Fig. 4A). Fault-slip data collected in this study are provided in supplementary Table S1.1

| GSA Data Repository Item 2012331, Table S1: fault-slip data, is available at www.geosociety.org/pubs/ftp2012.htm, or on request from editing@geosociety.org, Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA. |
Relative Timing and Characteristics of Faulting

Faults along the topographic break are highly variable in both geometry and kinematics. To aid in the identification of discrete fault populations associated with a particular stress tensor, the faults were assembled into groups based on crosscutting relationships and the nature of fault-zone material following procedures recommended by Angelier (1984), Liesa and Lisle (2004), and Sperner and Zweigel (2010). This led to the identification of three phases of faulting. The older phase I and II faults, which are referred to as early-stage faults, are associated with fault-zone material composed of quartz slickenfibers. The younger phase III faults, which are referred to as late-stage faults, are associated with fault-zone material composed of incohesive breccia and gouge.

The first phase of faults consists of 23 faults with surfaces laminated by quartz slickenfibers. These faults formed preferentially on bedding contacts between sandstone and slate horizons. Slickenfibers trend approximately downdip and give a thrust slip sense, which is consistent with flexural-slip folding. The second phase of faulting consists of a smaller group of 13 faults that...
crosstrack the bedding-parallel faults and postdate cleavage formation. These faults also contain fault-zone material composed of quartz slickenfibers, but they are oriented at high angles to bedding and are primarily strike-slip faults.

The third phase of faulting is composed of a population of 30 faults with distinctive fault-plane orientations and kinematics. A significant proportion of the late-stage faults are high-angle reverse faults that dip steeply toward the northeast. Crosscutting relationships indicate that late-stage faults consistently cut early-stage faults. In contrast to the early-stage faults, many of the late-stage fault zones are stained with iron precipitates, presumably because of permeability changes associated with faulting. The highest concentration of these faults occurs at field sites near the Peikang River.

The thickness of the fault-zone material associated with the late-stage faults ranges from centimeters to meters. One of the best examples of a relatively thick fault zone is located on a tributary of the Peikang River (Fig. 6B). The fault contains a 4-m-wide breccia zone, although most of the deformation appears concentrated within a 50-cm-wide gouge zone adjacent to the hanging wall. The fault, which strikes roughly 280° and dips 70° toward the NNE, accommodated at a minimum tens of meters of displacement. Riedel shears indicate a reverse slip sense, and pencil cleavage within a slaty layer in the hanging wall is dragged in an orientation consistent with reverse motion. The fault is located on the mapped trace of the Meiyuan fault, which forms the contact between the lower Tungmou Member and middle Lileng Member of the Paileng Formation. However, the fault strikes nearly perpendicular to the NE-trending Meiyuan fault and is kinematically incompatible with NW-SE compression.

Phase I and II faults are similar to faults previously described in the core of the Hsüehshan Range. Tillman and Byrne (1995) recognized flexural-slip faults similar to those observed in the phase I fault population. In addition, phase II faults have similar kinematics and crosscutting relationships to faults associated with a late phase of conjugate strike-slip faulting also previously recognized by Tilmann and Byrne (1995). Phase III faults, however, have not been recognized in the core of the Hsüehshan Range.

Inversion of Fault-Slip Data

The primary aim of paleostress inversion is to determine the orientations and relative magnitudes of the maximum (σ1), intermediate (σ2), and minimum (σ3) principal stresses or “paleostress axes.” These three principal stresses form the axes of the stress ellipsoid and represent the eigenvectors of some best-fit stress tensor that activated a fault population. In this study, we analyzed the fault-slip data using three inversion methods in order to compare the consistency of results, and we briefly discuss these methods here. For a comprehensive review of stress inversion techniques, see Angelier (1994) and Liesa and Lisle (2004).

P-T Axes Method and the Right Dihedra Method

In the strictest sense, the P-T axes method is not considered a stress inversion technique because the stress ratio $\Phi = (\sigma_1 - \sigma_3)/\sigma_1$, which describes the stress state during faulting, is not calculated. Rather, the P-T axes method is a kinematic approach that plots the orientations of the incremental principal shortening and

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**Figure 6. Photographs of late-stage faults.** (A) Minor reverse faults near Peikang River. Photograph is parallel to subvertical bedding plane. Note offset ripple marks. (B) Major reverse fault near tributary of Peikang River. Inset shows drag folding of pencil cleavage in hanging wall. Equal-area, lower-hemisphere stereographic projections in A and B show fault-slip data for photographed and nearby faults.
extension directions of each fault. The average P and T axes can then be calculated using a Bingham “unweighted” moment tensor summation, which is graphically analogous to contouring P and T axes (Marrett and Allmendinger, 1990). The calculated fault-plane solution represents the average kinematic axes.

The right dihedral method (Angelier and Meecher, 1977) provides a quick graphical technique to visualize the compressional and extensional quadrants of a fault population. This is accomplished by plotting shaded P and T quadrants repeatedly on top of one another for individual faults. The areas of greatest overlap presumably correspond to the areas where $\sigma_1$ and $\sigma_3$ plot. This technique is graphically analogous to the P-T axes method but better depicts the shape of the shortening and extension fields in stereographic projection. Once again, the stress ratio $\Phi$ is not constrained and is implicitly allowed to vary from 0 to 1 for each fault (Angelier, 1979, 1984).

**Gauss Paleostress Method**

Expanding on earlier direct inversion and grid-search stress inversion techniques that quantify the stress ratio $\Phi$, the functions embedded in the Gauss method allow the user to analyze the distribution of angular misfits ($\alpha$) from some best-fit stress tensor and separate polyphase fault populations by manipulating several user-defined parameters ($x, \Delta, \phi_s, \phi_f, k$) (Zalohar and Vrabec, 2007). In addition, as part of a compatibility measure, the Gauss method considers the mechanical stability of solutions by plotting the position of the “Mohr point” for each fault on a Mohr diagram. This process aids in constraining the best-fit stress tensor to mechanically realistic solutions for which Mohr points lie between $\phi_g$ and $\phi_f$. Thus, for every fault, the optimal stress tensor resolves acceptable normal to shear stress ratios consistent with the principles of Mohr-Coulomb failure. For a detailed discussion on the Gauss method, see Zalohar and Vrabec (2007).

The strength of the Gauss method lies in its ability to isolate homogeneous subsystems within a population of heterogeneous fault-slip data. Provided the angular differences between the principal stresses are $\geq25^\circ$, the Gauss method can accurately separate stress tensors, even when deviations in the angular misfit $\alpha$ are moderately large (e.g., $\leq55^\circ$). However, inhomogeneity in the stress field during faulting coupled with unevenly sized subpopulations can produce errors in the computation of the best-fit stress tensor (Zalohar and Vrabec, 2007). In order to eliminate the potential for hybrid solutions, fault-slip data were grouped into the previously described early- and late-stage subsystems before inversion. The Gauss method was then used to identify any misfit faults within each homogeneous subsystem and to help constrain the best-fit stress tensor to mechanically acceptable solutions.

**Paleostress Results Summary**

Spatial variations in fault kinematics were assessed by breaking the fault-slip data into groups based on their respective structural stations. For this analysis, the Gauss paleostress method was included because the threshold number of faults necessary to accurately calculate a reduced stress tensor was not exceeded at the Daan and Mei River structural stations. For each structural station, the results of the P-T axes and the right dihedral methods are presented in Figure 7, and eigenvectors calculated using the P-T axes method are given in Table 1. Faults were then grouped into early- and late-stage subsystems and analyzed using all three methods as depicted in Figure 8; the results of the Gauss paleostress method are given in Table 2. The six normal faults were removed, leaving 66 faults for stress inversion.

**Structural Stations**

1. **Daan River.** Two faults were identified, both of which are late-stage, oblique-slip back thrusters that strike parallel to mapped stratigraphic units. The calculated kinematic axes yield a shortening direction of NE-SW.

2. **Tachia River.** The data consist of 14 early-stage faults and six late-stage, oblique-slip reverse faults. The intermediate strain axis plunges gently toward the NNE because of the interplay between thrust and strike-slip faults. The calculated kinematic axes yield a maximum shortening direction of NW-SE.

3. **Peikang River.** In total, 23 faults were identified; nine are early-stage faults, and 14 are late-stage faults. Many of the late-stage faults dip steeply toward the NNE. The calculated kinematic axes yield a maximum shortening direction of NNE-SSW.

4. **Mei River.** The data consist of four early-stage faults and one late-stage fault. The calculated kinematic axes yield a maximum shortening direction of ENE-WSW.

**Early- and Late-Stage Subsystems**

**Early-stage subsystem.** The early-stage fault population consists of 36 faults. The maximum shortening direction calculated using the P-T axes method trends NW-SE, whereas the maximum shortening direction calculated using the right dihedral method lies in an ENE-WSW orientation. Gaussian stress inversion results are nearly identical with a maximum horizontal compression direction ($\sigma_{max}$) of $315^\circ$. In total, five faults were incompatible with the best-fit stress tensor.

**Late-stage subsystem.** The late-stage fault population consists of 30 faults. The maximum shortening direction calculated using the P-T axes method trends NE-SW; whereas the maximum shortening direction calculated using the right dihedral method lies in an ENE-WSW orientation. Gaussian stress inversion yields $\sigma_{max}$ of $259^\circ$, similar to the results of the right dihedral method. In total, three faults were incompatible with the best-fit stress tensor.

**Early-stage misfits.** For the early-stage subsystem, the five misfit faults were reanalyzed using the Gauss method, and a best-fit stress tensor was calculated. Stress inversion yields $\sigma_{max}$ of $243^\circ$. Based on kinematic consistency, we suggest that these faults were misidentified in the field and belong to the late-stage subsystem.

**Late-stage misfits.** For the late-stage subsystem, the three misfit faults were reanalyzed using the Gauss method, and a best-fit stress tensor was calculated. Stress inversion yields $\sigma_{max}$ of $350^\circ$, which is generally consistent with the phase II strike-slip faults noted at the core of the Hsuehshan Range (e.g., Tillman and Byrne, 1995; Fisher et al., 2002), suggesting these faults also may have been misidentified in the field.

**Timing of Regional Tilting**

Here, we briefly evaluate the timing relationship between the early- and late-stage fault populations and the tilting of strata. To accomplish this, the plunge of folds was rotated back to horizontal as observed at the core of the Hsuehshan Range. Faults were first rotated $26^\circ$ clockwise viewed to the northwest, and a best-fit stress tensor was then calculated using the Gauss method (Fig. 9). For the early-stage population, the $\sigma_{max}$ direction changed by only $5^\circ$, and six misfit faults were identified, five of which were the same as those identified using the original unrotated data. In addition, the misfit faults yielded an ENE-WSW $\sigma_{max}$ ($242^\circ$) analogous to previous results. Furthermore, removal of the axial plunge from the 23 phase I bedding-parallel
faults rotated the approximately downdip slip vectors into dip-slip orientations, consistent with flexural slip during folding. At a minimum, bedding-parallel faults within the early-stage fault population were affected by the regional-scale tilting. Irrespective, rotational effects did not substantially alter the early-stage stress inversion results presented in Figure 8.

Late-stage faults were also untilted around the same rotational axis. The majority of the NNE-dipping reverse faults were rotated into subvertical orientations, with their slip vectors plotting close to the center of the stereonet, indicating dip-slip motion. Based on the unreasonable fault-plane attitudes and kinematics, we suggest that the late-stage faults were not reoriented by regional tilting.

Cleavage measurements were also untilted; however, because the average pole to cleavage is approximately parallel to the rotation axis, and the rotation magnitude is only 26°, cleavage orientations were not significantly altered along the topographic break by the effects of tilting. As a result, cleavage still dips steeply toward the southeast, similar to data previously published from the core of the Hsüehshan Range (Tillman and Byrne, 1995). On the basis of these geometric arguments, we speculate that tilting of the southwest Hsüehshan Range occurred during the same period as the formation of the late-stage fault population, thereby rotating all of the older structures within the field area.

### SPATIAL VARIATION IN ROCK UPLIFT RATES FROM STREAM-PROFILE ANALYSIS

Topographic relief, elevation, and hillslope angles markedly change along the southwest flank of the Hsüehshan Range, which rises to >2500 m within 10 km of the Puli basin. This increase in elevation is accompanied by a change in hillslope angles, which average 32°–36° within the Hsüehshan Range and are typically less than 28° in the Puli topographic embayment (Chen et al., 2005). In tectonically active settings, steep slopes generally accompany high rock uplift rates. However, it is not possible to extract rock uplift rates directly from hillslope gradients in areas where uplift rates are high, as hillslopes generally have already reached slope stability thresholds (Binnie et al., 2007; Ouimet et al., 2009). Bedrock channel networks, in contrast, are argued to be in more direct connection with tectonic disturbances and may therefore be used to understand spatial variability in rock uplift rates even in high-uplift-rate environments (>0.5 mm/yr; Wobus et al., 2006; Ouimet et al., 2009; DiBiase et al., 2010). In this section, we extract longitudinal stream profiles...
from streams in the watersheds of the Tachia, Peikang, and Mei Rivers, which cut through the NNW-trending topographic break at a nearly perpendicular angle, and we argue that differences in channel profile steepness can be used to infer spatial variations in rock uplift rate across the mountain front.

**Theoretical Background**

Bedrock rivers are known to exhibit a scaled relationship between local stream gradient and contributing drainage area (Duvall et al., 2004; Wobus et al., 2006), which can be written as the power-law function:

\[ S = k_A^{-\theta} \]

(1)

where \( S \) is local stream gradient, \( A \) is contributing drainage area, and \( k \) and \( \theta \) are referred to as the steepness and concavity indices, respectively (Wobus et al., 2006). Theoretical work suggests that \( \theta \) should commonly fall within the range of 0.3–0.6 and be independent of rock uplift rate (Whipple and Tucker, 1999). For steady-state river profiles where rock uplift is balanced by bedrock incision, the relationship between channel steepness \( k \) and rock uplift rate \( U \) can be described by the following equation:

\[ k = \left( \frac{U}{K} \right)^{1/\theta} \]

(2)

where \( K \) is erodibility coefficient, \( U \) is rock uplift rate, and \( n \) is a coefficient that reflects the mechanism of incision and is \( \sim 1 \) for the detachment-limited stream power model (Whipple and Tucker, 1999; Wobus et al., 2006). This equation indicates a direct correlation between \( k \) and \( U \), provided that basin-scale lithology and climate, which affect \( K \), remain approximately uniform (e.g., Snyder et al., 2000; Kirby and Whipple, 2001; Kirby et al., 2003). Other studies, however, have demonstrated more complex relationships among \( k \), \( K \), and \( U \) due to the effect of variable rock strength, erosion thresholds, and the influence that channel narrowing can have on boundary shear stress conditions (e.g., Snyder et al., 2003; Duvall et al., 2004; Finnegan et al., 2005; Yanites et al., 2010). These studies indicate that variability in the erodibility coefficient and other nonlinearities in incision processes need to be addressed before uplift rates can be directly quantified from channel steepness indices.

In order to compare channel steepness across catchments of varying drainage area, steepness indices are typically “normalized” (\( k_{\text{ref}} \)) using a reference concavity \( \theta_{\text{ref}} \) (Wobus et al., 2006). Many studies assume values of 0.45 or 0.5 for \( \theta_{\text{ref}} \), which is well within the expected theoretical range of \( \theta \) (e.g., Wobus et al., 2006; Ouimet et al., 2009; DiBiase et al., 2010). Another approach is to set \( \theta_{\text{ref}} \) by calculating a mean concavity via log-S–log-A regressions of undisturbed stream profiles for a given field area (Duvall et al., 2004). For bedrock river profiles, the bounds of this regression should be fixed between a critical drainage area \( A_{\text{cr}} \), which defines the transition between hillslopes and channels (e.g., \( \geq 10^4 \text{ m}^2 \)), and a maximum drainage area \( A_{\text{max}} \), which coincides with a transition from bedrock to alluvial channel behavior that
were set to 106 m$^2$ and 109 m$^2$, respectively. The low, transitional, and high steepness color breaks were used to delineate zones of point data, and color classes were binned based on their distance from bedrock reaches. A kriging surface was generated from a DEM derived from aerial photographs. A batch profiler script was used to extract topographic data, and raw river profile data were smoothed over a 500 m moving window. Reference concavity gradients were calculated at 12 m contour intervals. Reference concavity $\theta$ was set to 0.57 based on a low-transformed slope-area regression of profile data extracted from both the Mei and Peikang Rivers, where $A_0$ and $A_{max}$ were set to 100 m$^2$ and 1000 m$^2$, respectively. The Tachia River concavity was not used to set the reference concavity value because major reservoirs are present along its length.

Normalized steepness indices ($k_n$) were calculated using a MATLAB script and ArcMap interface developed by Whipple et al. (2007) and made publicly available at www.geomorphtools.org. Topographic data were obtained from a Taiwanese 40 m digital elevation model (DEM) derived from aerial photographs. A batch profiler script was used to calculate $k_n$ for 0.5 km stream segments on all streams with contributing drainage areas $\geq 10$ km$^2$. Raw river profile data were smoothed over a 500 m moving window, and channel gradients were calculated at 12 m contour intervals. Reference concavity $\theta_{ref}$ was set to 0.57 based on a low-transformed slope-area regression of profile data extracted from both the Mei and Peikang Rivers, where $A_0$ and $A_{max}$ were set to 100 m$^2$ and 1000 m$^2$, respectively. The Tachia River concavity was not used to set the reference concavity value because major reservoirs are present along its length.

Normalized steepness indices calculated using the batch profiler script were imported into ArcMap and converted into point data. Reservoirs were identified using satellite imagery, and $k_n$ values from these channel segments were removed. A kriging surface was generated from point data, and color classes were binned based on Jenks natural breaks optimization. Distinct color breaks were used to delineate zones of low, transitional, and high steepness.

Spatial Variation of Normalized Steepness Indices ($k_n$)

To the west of the Shuangtung-Hsiaomao fault, streams incise through Pliocene and younger conglomerates and alluvium in the fold-and-thrust belt (Fig. 10). These rocks are substantially weaker than the Eocene–Miocene metasedimentary rocks upstream (Dadson et al., 2003; Yanites et al., 2010); therefore, we restrict the interpretation of $k_n$ values to bedrock reaches east of the Shuangtung-Hsiaomao fault. In detail, the $k_n$ values to the east of the Shuangtung-Hsiaomao fault delineate three zones of normalized steepness indices ($k_n$): low ($k_n < 420$), transitional ($420 < k_n < 770$), and high ($k_n > 770$) (Fig. 10). The mean and median $k_n$ values of the over five-thousand 0.5-km-long stream segments within the entire catchment were calculated to be 640 and 450, respectively. Both the mean and median $k_n$ values fall within the range of transitional steepness values, suggesting that the low and high steepness zones accurately depict relatively low and relatively high $k_n$ values.

High $k_n$ values observed along the Tachia and Peikang Rivers as they flow out of the Hsüehshan Range and in the region west of the Shuangtung-Hsiaomao fault were excluded in the spatial delineation of the low, transitional, and high steepness zones. These locally high $k_n$ values appear out of place given the lower $k_n$ values found within the adjacent tributaries, and are coincident with wider channels, large boulder deposits from floods and/or debris flows, and the general alluvial character of these downstream trunk river reaches.

In general, the observed distribution of $k_n$ and our generalized zones do not spatially correlate with faults mapped in the Hsüehshan Range. This is illustrated by the transitional steepness zone, which forms a recess curving around the Puli topographic embayment, mirroring the map pattern of the 1000 m contour. Observed $k_n$ values increase by up to one order of magnitude across this transition zone, which, to the north of the Puli basin, is parallel to the NNW-trending topographic break. For example, upstream of the Meiyan fault, steepness indices abruptly climb to >1500 and reach nearly 4000 just 2.5 km upstream. In the following sections, we investigate whether this pattern of $k_n$ values is related to variations in river behavior and physical processes that limit the rate of incision or to variations in uplift rate.

Influence of Lithology and Sediment Transport on Steepness Values

Rock strength can significantly impact channel steepness (e.g., Duvall et al., 2004). Schmidt hammer rebound and joint spacing measurements collected along the Peikang River indicate that rock strength is highly variable at the outcrop scale (Yanites et al., 2010). Despite this localized variability, Yanites et al. (2010) concluded that over larger distances, there is no statistically significant change in rock strength within the metasedimentary Eocene–Miocene strata. Although previous studies have shown a modest spatial correlation between compressive rock strength and average hillslope angles islandwide ($r^2 = 0.31$; Dadson et al., 2003), the elevated hillslope angles in the Hsüehshan Range are more likely a response to tectonic forcing rather than a substantive change in compressive rock strength. Outcrop-scale variations in rock strength may be responsible for generating fluctuating $k_n$ values, as observed within the high steepness zone. Equally so, this variability may be caused by valley floors approaching the resolution of DEM pixels (40 m) in narrow channel segments (Yanites et al., 2010) or by propagating knickpoints and transient landslide dams. Regardless, the transitional steepness zone does not correlate with mapped lithologic boundaries, indicating that bedrock lithology and rock strength are not important factors controlling the spatial distribution of $k_n$ values within the field area.

Bedrock river behavior, which can vary from detachment- to transport-limited end members, may also affect the morphology of rivers and interpretation of channel steepness (Whipple, 2004). The detachment-limited model is widely used, and it argues that channel incision and...
adjustment are driven by stream power, which controls the rivers capacity to detach bedrock particles (e.g., Whipple and Tucker, 1999). The transport-limited model, in contrast, argues for the importance of sediment transport and supply, specifically that channel incision and adjustment are driven by downstream sediment flux variation (e.g., Whipple and Tucker, 2002). Rivers with small drainage areas, those incising into strong rocks, and/or those dissecting landscapes exhibiting high rock uplift rates are typically argued to be governed by detachment-limited behavior (Whipple, 2004). Yanites et al. (2011) inferred that the Peikang River behaves more closely to the transport-limited end member because channel width influences shear stress more strongly than slope. Although this may be the case for certain portions of the trunk rivers in the watersheds analyzed for this study, we note that a similar distribution of $k_{sn}$ values is observed in the side tributary network regardless of whether or not trunk rivers are included in the interpolation. These steep, smaller side tributaries are much more likely to be governed by detachment-limited behavior, and thus we argue that the overall behavior of the fluvial network more closely approximates that of detachment-limited bedrock rivers.

**Integration with Precise Leveling Data**

Recently published geodetic leveling data were used to determine if the pattern of $k_{sn}$ values correlates with independent measurements of rock uplift rates. The Taiwan vertical datum (TWVD2001) is a network of 1843 benchmarks anchored in bedrock along public roads and spaced ~2 km apart. From 2000 to 2008, four leveling campaigns were conducted to measure islandwide vertical displacement rates (Ching et al., 2011; K.H. Chen et al., 2011). The mean error for islandwide data is ±1.64 mm/yr (K.H. Chen et al., 2011), making this data set the highest-precision vertical-displacement-rate field currently available for Taiwan.

An east-west profile of rock uplift rates across the low, transitional, and high steepness zones was constructed from 14 leveling benchmarks (Fig. 11). The leveling line begins in the Puli basin, follows the Mei River upstream, and eventually crosses the Lishan fault. Five benchmarks are in the Puli basin in the low steepness zone, five are in the transitional and high steepness zones, and four are to the east of the Lishan fault in the western Backbone Range. Uplift rates increase from 5.15 ± 1.55 mm/yr in the Puli basin to 15.9 ± 1.88 mm/yr just west of the Lishan fault. This differential uplift rate of 10.75 mm/yr within a distance of ~11 km is similar to that calculated by Yanites et al. (2010) using a calibrated river incision model. At least within the vicinity of the Mei River, this increase in uplift rate coincides with the transitional steepness zone, indicating that the $k_{sn}$ values broadly reflect changes in rock uplift rate.

The agreement between geodetic rock uplift rates and the spatial distribution of $k_{sn}$ values suggests that the contemporary uplift pattern has been long-standing over the Holocene, the period of time during which bedrock channels equilibrate with rock uplift rates. Recent observations by Yanites et al. (2010) from the Peikang River show that both slope and width change in response to increased rock uplift and accelerated incision. Specifically, the river will narrow in response to increases in rock uplift until a minimum width-depth ratio is achieved, at which point the river will begin steepening. The agreement between geodetic rock uplift rates and $k_{sn}$ values suggests that sufficient time has passed for the channels to adequately narrow...
and for channel steepening to dominate the fluvial response in the Hsiehshan Range.

**DISCUSSION**

The preceding field observations and structural analyses indicate that the early- and late-stage fault populations formed within stress fields of different orientation, spatial extent, and relative age. The maximum compression direction for the early-stage fault population, which trends between 310° and 315°, depending on whether the tilted or untilted data are considered, is in broad agreement with the overall shortening direction of the collision zone in Taiwan and with mapped fault and fold traces in the field area. In addition, the geometry, kinematics, and age relation of the phase I and II faults within the early-stage fault population can be observed throughout the Hsiehshan Range (Clark et al., 1993; Tillman and Byrne, 1995; Fisher et al., 2002), indicating that the early-stage faults did not form in a stress field that was localized to the topographic break. The ubiquitous presence of quartz slickenfibers and lack of gouge and breccia on early-stage fault surfaces further suggest that the faults did not accommodate an appreciable amount of slip in the uppermost few kilometers of the crust. This implies that the early-stage faults have been inactive and that their exposure is a result of extended uplift and erosion. Based on these observations, we propose that the early-stage faults were formed during a period of NW-SE compression that deformed the entire package of Eocene–Oligocene strata in the Hsiehshan Range.

In contrast to the early-stage faults, the ENE-WSW maximum compression direction for the late-stage fault population lies ~60° from the typical maximum compression direction in Taiwan and is approximately normal to the 345°-trending topographic break. We propose that this kinematic uniformity with local topographic relief indicates that the Hsiehshan Range has been uplifted along late-stage faults. Furthermore, the late-stage faults have only been recognized within and near the kₖ transitional steepness zone, suggesting that these faults are related to local tectonic processes and that the transitional steepness zone encompasses a high concentration of these faults. Unlike the early-stage faults, the late-stage fault zones contain incohesive breccia and gouge, which are characteristic of relatively shallow depth during faulting. These observations lead us to conclude that the late-stage faults formed during relatively young and shallow brittle faulting that is localized to the area within the transitional steepness zone. In addition, a number of midcrustal earthquakes (~18 km depth) located beneath the northeastern border of the Puli basin (Wu and Rau, 1998) have focal mechanism solutions with ENE-WSW maximum compression directions. These earthquakes suggest that the tectonic processes driving the formation of the late-stage faults are ongoing.

The presence of early- and late-stage faults within the field area indicates that the stress field has varied significantly both spatially and temporally in central Taiwan. The stress field responsible for the generation of the late-stage faults is broadly consistent with the map pattern of kₖ values determined in this study, suggesting that the spatial uplift patterns are a reflection of slip occurring along late-stage faults. In the following section, we evaluate a number of other data sets that constrain uplift rates in the Puli basin area over a variety of temporal scales in order to understand the tectonic evolution of the region.

**Spatiotemporal Evolution of Rock Uplift Rates**

Late Quaternary to Holocene bedrock incision rates in the area near the topographic break were recently quantified along the Peikang River by Yanites et al. (2010). They found that incision rates in the footwall of the Meiyuan fault (~2–3 mm/yr) calculated by dating strath terrace deposits with optically stimulated luminescence (OSL) were relatively low compared to incision rates in the hanging wall (~9–13.5 mm/yr) calculated with a calibrated river incision model. Several kilometers downstream, Holocene incision rates determined from OSL data reach a maximum of ~6–10 mm/yr in the hanging wall of the Shuilikeng fault. Farther downstream, laterized soils and OSL samples indicate incision rates of ~2–4 mm/yr in the hanging wall of the Shuangtung-Hsiaomao fault. In short, if the Peikang River has approximated steady-state form over the Holocene, moving in the direction downstream, the data signify a dropoff in uplift rates across both the Meiyuan and Shuilikeng faults and a gradual increase in uplift rates between them.

In comparison, contemporary geodetic data in the vicinity of the Peikang River show a downstream decrease in rock uplift rates beginning near the Meiyuan fault (Ching et al., 2011). This is shown by the GPS station HUYS (green triangle in Fig. 10) in the hanging wall of the Meiyuan fault, which records a vertical velocity of 9.5 mm/yr (Ching et al., 2011), while ~6 km downstream, leveling benchmarks record a decrease to 4.5 mm/yr, and farther downstream reach 0.5 mm/yr near the Shuilikeng fault. However, if the number of leveling benchmarks analyzed is expanded to all of those west of the 1000 m contour and east of the Shuilikeng fault, present-day uplift rates average ~6 mm/yr in the Puli topographic embayment, which is broadly consistent with Holocene incision rates downstream of the Meiyuan fault. When leveling data from the Mei River are also included (Fig. 11), the geodetic data define a NNW-trending break in uplift rates. It is probable that this zone has been active since at least the Holocene, given the OSL incision patterns along the Peikang River and the similar boundary defined by the kₖ transitional steepness zone. Differential uplift across this NNW-trending zone was likely partially accommodated by the late-stage, WNW-striking, high-angle reverse fault
depicted in Figure 6B (yellow star in Fig. 10), rather than by the mapped NE-striking Mei-yuan fault.

Palynological and radiocarbon studies discussed by Chen et al. (2005) provide additional information on the evolution of rock uplift rates over the time span of hundreds of thousands of years in the Puli topographic embayment. Well-log data from sediments in the Yuchi, Sun Lake, Moon Lake, and Toushe basins indicate that the Puli basin chain formed as early as ca. 200 ka, which implies that uplift rates were subdued before the initial formation of the basins (Chen et al., 2005). Given the physiographic expression and metamorphic grade of the Hsüehshan Range, it appears that to the northeast of the topographic break, uplift rates over the same time period were higher than in the Puli topographic embayment. The differential uplift rate between the embayment and Hsüehshan Range along the Mei River (~11 mm/yr) could generate >2000 m of topographic relief over 200 k.y. and is consistent with the topographic relief relative to the Puli basin (~1000s m). Although difficult to quantify, the available data indicate that rock uplift rates in the Puli topographic embayment have been subdued relative to the Hsüehshan Range to the northeast since at least 200 ka and probably somewhat longer.

**Stress Trajectories around the Continental Margin Promontory**

The effect of the Peikang high on stress trajectories in western Taiwan was examined by Jeng et al. (1996) using a numerical finite-element model. Stress trajectories from their study were repositioned in Figure 12A toward the northeast so that the margin of the modeled indenter coincides with the Sanyi-Puli seismic zone, the heel-shaped seismogenic boundary (Lin, 2001), and the continental margin magnetic anomaly high (Byrne et al., 2011). This repositioning allows us to compare the maximum compression direction from the early- and late-stage fault populations to those simulated for a collision zone with a rigid indenter of similar geometry to the continental margin promontory. The modeled \( \sigma_1 \) trajectories, which exhibit a fan-shaped pattern and converge toward the indenter, are slightly oblique to the orthogonal of the Sanyi-Puli seismic zone and the continental margin promontory, with a \( \sigma_1 \) orientation associated with the late-stage fault population. In addition, the topographic break coincides with the modeled \( \sigma_{max} \), which lies 20° clockwise from the 325°-trending Sanyi-Puli seismic zone. Following this interpretation, we propose that the maximum compression direction of the late-stage faults is related to a local change in the orientation of the stress field caused by the underthrusting continental margin promontory and fracture zone.

**Tectonic Model of Central Taiwan**

Several authors have put forward models to explain the formation of the Puli topographic embayment (Simoes and Avouac, 2006; Wilcox et al., 2011); however, there is no current consensus on the mechanism and timing of formation of this zone of low relief and elevation. Here, we integrate the structural and geomorphic data presented in this study with previous thermokinematic modeling results and a number of other data sets and present a tectonic model that can explain the structural and topographic development of central Taiwan.

The most recent thermokinematic modeling results suggested that frontal accretion of the Eurasian continental crust can explain the observed deformation and exhumation patterns in Taiwan (Yamato et al., 2009). However, other recent papers have proposed that 50%-90% of the material influx into the orogenic wedge during the mature phase of the collision is accommodated through underplating rather than frontal accretion (Fuller et al., 2006; Beyssac et al., 2007; Simoes et al., 2007). Simoes et al. (2007) proposed that separate underplating windows beneath the Taranui metamorphic complex and Hsüehshan belt can reproduce the islandwide spatial distribution of cooling ages determined from a variety of thermochronometers. In the Hsüehshan belt, underplating may have begun as early as 4 Ma and from 1.5 Ma to present became the dominant uplift mechanism (Simoes et al., 2007). The underplating window beneath the Hsüehshan belt is estimated to be between ~10 km and ~18 km in depth and ~15 km in width (Simoes et al., 2007), and its western border lies near the Shuilikeng fault.

The depth of this modeled underplating zone coincides with the depth of the ENE-WSW compression earthquakes within the Sanyi-Puli seismic zone (Wu and Rau, 1998) and is at a similar depth to the triangular zone of mechanically strong crust revealed by \( Q_P \) and \( Q_S \) seismic attenuation data south of the seismic zone (Wang et al., 2010). At 19 km depth, the eastern margin of this relatively strong crust follows the trend of the magnetic anomaly high, suggesting these data are imaging the geometry of the collided promontory (Byrne et al., 2011).

As previously discussed, the formation of the Puli basin chain at least by 200 ka is the first definitive indicator that uplift rates were subdued in central Taiwan. However, it is probable that uplift rates were relatively low for a substantial period of time before the formation of the basins. For example, peak metamorphic temperatures attained by rocks surrounding the Puli topographic embayment are lower than those at the core of the Hsüehshan Range. Specifically, the 330–350 °C Raman spectroscopy of carbonaceous materials (RSCM) temperature zone (Beyssac et al., 2007) bends around the continental margin promontory (Fig. 12A). This is consistent with the inferred offset in illite crystallinity from earlier studies (Fig. 3; Chen et al., 1983). Given the inferred offset in metamorphic grade, we suggest that uplift rates in the Puli topographic embayment have been subdued significantly longer than 200 k.y. This interpretation is consistent with Wilcox et al. (2011), who suggested that the Puli topographic embayment may have initiated between 500 ka and 700 ka, as well as Chen et al.’s (2005) inference that uplift rates slowed before the Puli basin chain formed.

The tectonic model presented in Figure 12 integrates the data discussed in this contribution with the collision of a continental margin promontory originally proposed by Byrne et al. (2011). Byrne et al. (2011) proposed that the continental margin promontory resulted in a threefold division of the fold-and-thrust belt as it developed from ca. 2 Ma to present. In southern Taiwan, continental crust of normal thickness (~30 km) limited westward propagation of the thrust belt, which may have initiated lateral extrusion in southern Taiwan. In the north, thick sedimentary sequences of the Taishi basin allowed the thrust belt to propagate substantially west of the promontory, resulting in a left-lateral offset between the northern and southern belts. In central Taiwan, in the area of the promontory, progressively younger thrusts rotated counterclockwise as the northern belt advanced and the southern belt remained pinned by continental crust.

Our model builds upon this interpretation and consists of two stages: an early stage that involved the subduction of transitional crust of variable thickness during the propagation of the fold-and-thrust belt and a late stage that began with the collision of crust of normal thickness, i.e., the continental margin promontory, with a zone of underplating beneath the Hsüehshan belt. By extrapolating an underthrusting rate of ~42 mm/yr relative to the Longitudinal Valley fault (Simoes and Avouac, 2006) in the direction of relative plate convergence, we find that the continental margin promontory would have begun to subdue beneath the zone of underplating around 0.5 Ma, defining, for the purpose of our two-stage model, the beginning of the late-stage collision. In the context of this model, the Puli topographic embayment formed later than the initial propagation of the fold-and-thrust belt.

During the early-stage collision, maintenance of foreland taper to the west of the Shuilikeng
Nonparallel topographic and structural curvature in an active collision zone

Partial subduction of continental margin promontory

Shuangtung-Hsiaomao Fault

Peikang Basement High

Present

Changhua Fault

Shuilikeng Fault

Paimaoshan Fault

Chelungpu Fault

Late-stage faults

Puli basin

Collision with promontory

Underplating beneath central Hsüehshan belt stops

Partial subduction of continental margin promontory

Late-stage faults

σ1

σhmin

Topographic Break

~20˚

RSCM Temperature

< 330° C

330 - 350° C

350 - 450° C

450 - 500° C

0 20 40 60

km

Figure 12. Tectonic interpretation of structural development of southwestern Hsüehshan Range and Puli topographic embayment. (A) Map of Taiwan showing Peikang basement high, continental margin promontory as defined by magnetic anomaly high and Sanyi-Puli seismic zone, topographic break, modeled stress trajectories, Raman spectroscopy of carbonaceous materials (RSCM) peak metamorphic temperatures, and orientation of σ1 determined from late-stage fault population. Stress trajectories were modeled for Peikang basement high by Jeng et al. (1996) but are shifted to the east to correspond to inferred position of continental margin promontory. RSCM peak metamorphic temperature map is modified from Beyssac et al. (2007). (B–C) Collision of continental margin promontory at ca. 0.5 Ma with zone of underplating beneath Hsüehshan belt resulting in reactivation of Sanyi-Puli seismic zone, decrease in rock uplift rates in region of Puli topographic embayment, and development of Paimaoshan fault, late-stage faults, and SW-plunging folds along topographic break.
fault was accommodated through frontal accretion, while in the Hsiüehshan Range, underplating had already begun uplifting the mountain belt (Simoes et al., 2007; Beyssac et al., 2007). At shallow structural levels, thin-skinned foreland thrust faults propagated westward, becoming progressively more arcuate because of the increasing interaction with the continental margin promontory (Fig. 12B). At deeper structural levels, basement rocks that comprise the continental margin promontory were underthrust beneath the toe of the décollement without disrupting the frontal accretion process. The early-stage faults observed in the core of the Hsiüehshan Range and along the topographic break formed during this stage of the collision, when the maximum compression direction approximately paralleled the relative plate convergence vector.

During the late-stage collision, the relatively strong continental margin promontory collided with duplexes underplating the Hsiüehshan belt at ca. 0.5 Ma (Fig. 12B) and was subsequently partially subducted (Fig. 12C). As the leading edge of the continental margin promontory was thrust beneath the region of underplating, the relatively strong footwall indenter may have impeded the formation of duplexes, thereby slowing vertical uplift rates in the area of the Puli basin. Faults to the west in the fold-and-thrust belt, such as the Changhua and Chelungpu faults, however, remained active. In addition, uplift rates in the Hsiüehshan Range were not affected because the region of slowed underplating was spatially controlled by the geometry of the promontory. Over the past ~0.5 m.y., the differential uplift of the Hsiüehshan Range relative to the Puli topographic embayment was accommodated by the late-stage faults. The higher uplift rates in the Hsiüehshan Range resulted in the tilting of folds toward the southwest and the formation of the topographic break in a NNW-trending orientation consistent with the stress field related to the underthrusting promontory and fracture zone.

A cluster of 20- to 25-km-deep earthquakes appears to wrap around the leading edge of the promontory (Wu et al., 2009), possibly highlighting the collision of the promontory with duplexes beneath the Hsiüehshan Range. We suggest that as the promontory became involved in crustal shortening, the continental margin fracture zone reactivated as a left-lateral transfer fault, which is presently illuminated by earthquakes within the Sanyi-Puli seismic zone. Further evidence for this reactivation is shown by the Paimaoshan fault, which is located just to the southwest of the transitional steepness zone and directly above the Sanyi-Puli seismic zone (Figs. 4 and 12C). This 10-km-long fault follows the trend of the Sanyi-Puli seismic zone and is younger than the Shuilikeng fault, which it offsets in a left-lateral sense.

Before the collision of the promontory, the topographic grain of the Hsiüehshan Range to the east of the Shuilikeng fault likely followed the structural trend of faults and folds in the Puli area. As the collision of the promontory slowed uplift and erosion rates, mass-wasting processes progressively lowered valley slope angles in the Puli region. These processes were strongly influenced by the glacial climates that produced less-vegetated, unstable hillslopes over the past 200 k.y. (Chen et al., 2005). For example, lithofacies studies from sediments in the Puli basin chain indicate that infrequent but catastrophic rainfalls evacuated large amounts of sediment surrounding the Puli area via landsliding and debris flows (Chen et al., 2005). Provided that fluvial incision rates slowed in tandem with the decrease in rock uplift rates, these processes over ~0.5 m.y. could adequately lower hillslope angles, forming an area of relatively low relief as the rocks north of the promontory continued to be uplifted.

CONCLUSIONS

The results presented in this study indicate that the Eurasian continental margin architecture exerts a fundamental control on the three dimensions of mountain building in Taiwan. In this study, two populations of faults, an early stage and a late stage, were identified along the south flank of the Hsiüehshan Range. Paleostress results yield a maximum horizontal compression direction that trends NW-SE for early-stage faults and ENE-WSW for late-stage faults. We propose that the early-stage faults formed during the initial thin-skinned development of the Hsiüehshan belt and that the late-stage faults formed in response to the continental margin promontory colliding with duplexes underplating the Hsiüehshan belt. In addition, a NW-trending break in the primary values coincides with the topographic break, an increase in geodetic uplift rates, and a high concentration of late-stage faults, suggesting that the Hsiüehshan Range is uplifting faster than the Puli topographic embayment along the late-stage faults. The k values from this study in conjunction with an array of other data sets indicate that rock uplift rates have been subdued in the Puli topographic embayment relative to the Hsiüehshan Range since at least ca. 0.5 Ma, when the continental margin promontory is interpreted to have begun collision. The decrease in uplift rates relative to the surrounding mountains led to the formation of the Puli topographic embayment. Furthermore, based on focal mechanism solutions from the Sanyi-Puli seismic zone and the displacement of the Shuilikeng fault by the Paimaoshan fault, we propose that the continental margin fracture zone has been reactivated as a left-lateral transfer fault since the late-stage collision of the promontory. This tectonic model argues that topography in Taiwan is sensitive over relatively short time scales (hundreds of thousands of years) to tectonic forcing and that the topographic grain of active mountain belts responds to variations in the thickness, composition, and geometry of crust in the downwelling plate more quickly than region-scale, through-going, structural features.

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