QUATERNARY DEFORMATION ALONG THE NORTH WUTAISHAN FAULT IN THE SHANXI RIFT SYSTEM, CHINA

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by
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The undersigned, appointed by the dean of the Graduate School, have examined the thesis entitled:

Quaternary Deformation along the North Wutaishan Fault in the Shanxi Rift System, China

Presented by John H. Corley
A candidate for the degree of Master of Science
And hereby certify, in their opinion, it is worthy of acceptance.

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Dr. Francisco Gomez

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Dr. Mian Liu

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Dr. Michael Urban
This thesis is dedicated to all of my family, friends—old and new, near and abroad—and teachers who have helped me get to this point. Without their constant support (and prodding), this thesis would have been much more difficult to finish and way less fun.

Thank you.
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QUATERNARY DEFORMATION ALONG THE NORTH WUTAISHAN FAULT IN THE SHANXI RIFT SYSTEM, CHINA

John H. Corley

Dr. Francisco Gomez, Thesis Supervisor

ABSTRACT

The Shanxi Rift System in northern China is a tectonically active area in an intraplate setting that consists of northeast-southwest oriented half-graben basins. This study used data from multiple scales to investigate the influence of tectonics on features ranging in size from the meso- to regional scale. The data utilized include fault kinematic indicators, stream terraces, and regional/basin-wide morphometry data. It was found that paleostream profiles created from a stream's terraces generally follow the modern stream profile, however, there was evidence of converging profiles in the upper reaches of the Yangyan River. This convergence may be a result of fault block tilting. Offset stream terraces found on either side of a smaller fault splaying off of the Wutaishan Fault were calculated to have fault throw/uplift rates ranging from 0.29-0.58 mm/yr. This rate is much less than the Wutaishan fault rate of 1.55-2.00 mm/yr, so this suggests that there is a spatial variation of faulting rates and strain accumulation in the area. Quaternary faulting rates were also calculated for the Wutaishan fault from terrace height and age data, and it was found that from about 1.2 Ma to 0.13 Ma, the fault throw/uplift rate was relatively constant at 0.22 mm/yr to 0.41 mm/yr. Then, sometime around or after 0.13 Ma, the rate greatly increased to a range of 0.64-1.07 mm/yr. Geomorphic indices of morphometric analyses were used to assess the tectonic activity in the area. The results of Stream Length-Gradient Index and Hypsometric Integral analyses for Strahler orders 2, 3, and 4 watersheds show that the highest morphometry values were found near the
bend of the Wutaishan fault, towards the western end of the Daixian-Fanshi Basin. This suggests that there may be a correlation between higher morphometric indices and strain accumulation at a fault bend. An analysis of fault kinematic indicators found at an outcrop reveal NW-SE extension; this compares well to the overall structure of the Daixian-Fanshi Basin/Wutaishan fault. However, a reanalysis of 13 earthquakes in North China using a moment tensor summation reveal that the orientation of the P and T-axis of a fault plane solution is similar to a strike-slip or "wrench" fault. These two analyses suggest that the strain accumulation for the basins in the Shanxi Rift System is locally controlled by normal faults, and that the strain accumulation for region as a whole is accommodated by a “wrenching” type of deformation (i.e., a regional, horizontal shear) in response to tectonic forces.
Chapter 1 – Introduction and Research Questions

I. Introduction

The Shanxi Rift System (also known as the Shanxi Graben System) in northern China is a tectonically active area located within an intraplate setting and characterized by high amounts of seismicity throughout history (Figure 1.1). Volcanic arc and continental block collisions in the Archean formed what is now known as the North China Block (NCB), which is the amalgamation of two smaller blocks separated by a mountain belt: the Western China Block, Central Orogenic Belt, and the Eastern China Block (Kusky et al., 2007). Orientations of faults and basement fabrics from these past orogenies, in conjunction with forces related to continent-continent collision and subduction, have led to the pull-apart basins seen today (Xu et al., 1993a; Zhang et al., 1998). Faults in this area have produced many destructive earthquakes, some of which have killed hundreds of thousands of people. The recurrence intervals of earthquakes and patterns of seismicity are not completely understood for many areas in this rift system. Therefore, many studies have utilized paleoseismic, geomorphologic, and geophysical techniques to better understand the relationships between tectonics and geology.

II. Research Questions and Goals of the Study

The main goal of this study is to evaluate the influence of tectonic processes on the landscape in the Shanxi Rift System (SRS). More specifically, the effects of tectonic uplift and its influence on river systems and their terraces will be analyzed. Other aspects of this study will examine structural data and basin formation. Some questions that will be explored include:
Figure 1.1  Shaded relief map of the central and northern sections of the Shanxi Rift System. Inset indicates location in relation to Asia; star indicates location of the study area.
1) How does the current profile of a stream compare to the paleoprofile of the stream reconstructed from relict stream terraces?

2) Can spatial patterns and variability of regional uplift/subsidence be inferred from paleostream terrace measurements along a river?

3) What does the regional morphometry suggest for spatial patterns of tectonism?

4) Can variation in uplift/subsidence rates be inferred at a basin-wide scale? For example, does one side of the basin uplift/subside at a different rate than the other side, thus introducing a tilted block scenario?

5) What are the basin kinematics?

The first two questions regarding uplift rates were assessed using data collected from the field, such as paleoterrace elevation measurements, and also by performing an analysis of digital elevation model (DEM) data. DEM analysis also included the delineation of stream channels, which were used to create longitudinal river profiles. These generated profiles of the modern stream were compared to paleostream profiles that were created from paleoterrace height measurements. Further DEM analysis included calculating the Stream Length-Gradient Index (SL index) and hypsometric integrals for watersheds based on stream order. The fifth question regarding kinematic analyses in the Shanxi grabens was addressed using kinematic analysis of mesoscopic fault data.

The findings of this study provide a way of understanding the forces and resulting geometries of rifting in the interior of a continent. This area is seismically active, which is not common in an intracontinental setting; most earthquakes occur at plate boundaries. Other implications discussed will include debate of whether or not the Shanxi Rift is an
active or passive rift. For example, is the formation of the rift a result from tectonic collision or mantle upwelling? The results can help elucidate possible patterns in seismicity throughout time, which will lead to improving earthquake probability analysis. Research of the SRS's structural and seismological properties can supplement knowledge regarding similar intracontinental rifts such as the Baikal Rift or the Rio Grande Rift, or failed rifts in the case of the New Madrid Seismic Zone.
Chapter 2: Geologic Background of the Shanxi Rift System

Even though the Shanxi Rift System (SRS) began to develop in the later half of the Cenozoic, it is fundamental to understand that the current orientation of the rift can be attributed to forces acting upon pre-existing oriented structures and rock fabrics in the basement. These pre-existing basement features are the result of ancient orogenies that created zones of weak suturing between multiple continental blocks (Xu et al., 1993a; Zhang et al., 1998; Kusky, 2011).

I. Geology and Tectonics of the Precambrian through the Mesozoic

The SRS is located on the eastern end of the Eurasian plate's North China Block (NCB). More specifically, the SRS is located between two smaller continental blocks, which are delineated by numerous active faults (Figure 2.1). The NCB consists of two major continental blocks, the Eastern and Western blocks, separated by a Precambrian orogenic belt (Figure 2.2). The exact timing and style of formation of this orogenic belt between the Eastern and Western blocks is the subject of much debate (Zhao et al., 2001; Kusky, 2011). The primary contention is whether the orogeny that separates the two blocks formed in the Archean (Kusky, 2011) or Proterozoic (Zhao et al., 2001). Besides timing of collision, another point of debate is the orientation of subduction zones in the orogenic process (Zhao et al., 2001; Kusky, 2011).

The main focus of this thesis is in the region that separates the Eastern and Western Block of the NCB, termed the Central Orogenic Belt (COB) by Kusky (2007). In this area, the SRS would initiate during the Cenozoic due to forces acting on lithologic fabrics that were oriented from a previous orogeny.
Figure 2.1  Map of tectonic blocks and faults in the region surrounding the Shanxi Rift System. The star indicates study area. Figure modified from Kusky et al., (2007).
Figure 2.2  Geologic and tectonic subdivisions of the North China Craton. After subsequent orogenies and subduction events, basement structures and fabric orientation of the Central Orogenic Belt would play a major role in the influence of rifting in the Cenozoic. The study area is located within the Central Orogenic Belt. Figure modified from Kusky (2011).
The proposed formation of the COB occurred when the Eastern Block collided with an arc above a western-dipping subduction zone around 2.5 Ga, and then collided with the Columbia supercontinent to produce a large-scale metamorphic event at 1.85 Ga (Figure 2.3). In support of Kusky's model of westward-dipping subduction, Zheng et al., (2009) provides geophysical evidence of a westward-dipping subduction zone. Zheng used receiver functions to image the crust and upper mantle beneath the Eastern Block, COB, and the Western Block of the NCB. Zheng's results indicate two westward-dipping low velocity structures as possible evidence of paleosubduction beneath the COB.

The NCB was relatively stable from about 1.9 Ga to 250 Ma, after which the area was subjected to multiple orogenic events in addition to deformation related to a migrating subduction margin to the east of the NCB. There are two main orogenies in the Mesozoic: the Indosinian (Triassic-Jurassic) and Yanshansian (Jurassic-Cretaceous) orogeny (Kusky et al., 2007). These two orogenies rotated E-W oriented structures into NNE-trending fold and fault structures. It is the eventual NE-SW orientation of the structures and fabrics found within the COB (Figure 2.2) that sets the stage for Cenozoic rifting that preferentially initiates and propagates along these zones of weaknesses between the Eastern and Western Blocks (Xu, et al., 1993a, Zhang et al., 1998).

During this time, subduction of the Kula plate along the eastern margin of the NCB was active from 200-100 Ma (Kusky et al., 2007). The subducted oceanic crust dehydrated, which hydrated and weakened the overlying mantle underneath the Eastern Block. As a result, the lithosphere thinned and delaminated, potentially leading to the formation of the extensional tectonics in the NCB, such as the Shanxi Rift (Kusky et al., 2007, and references therein).
Figure 2.3  Model of the Precambrian evolution of the North China Craton. WB – Western Block, COB – Central Orogenic Belt, EB – Eastern Block. Note that the COB is formed around 2.5Ga, and then undergoes metamorphism at 1.8Ga. Figure modified from Kusky (2011).
II. Cenozoic Geology and Tectonics

The Western Block, which includes the Ordos Plateau, is a stable block of continental lithosphere that exhibits very low seismicity, low heat flow, thick mantle root, and strongly negative Bouguer anomalies. In contrast, the Eastern Block is the location of many destructive earthquakes, high heat flow, thin crust and lithosphere, low positive or weakly negative Bouguer anomalies, and 2-3 km of mantle upwarp in the southern part of the SRS (Li et al., 1998; Kusky et al., 2007; Lei, 2012; Yu et al., 2012).

The timing of formation of the SRS is mostly known from stratigraphic and paleontologic evidence. The oldest records of rift initiation are in the southernmost grabens of the SRS, mainly the Weihe Basin (also known as the Wei River Basin). According to Zhang et al., (1998, and references therein), lacustrine and alluvial deposits found in the Weihe Basin indicate that the graben first began to develop in the mid to late Eocene. These deposits are limited to a few troughs in the Weihe Basin, but up to 1500 m of lacustrine sediments are found. During the mid Miocene, rifting and deposition spread to the whole basin with up to 2500 m of sediment accumulation, and calculated subsidence rates of 0.20 mm/yr (Bellier et al., 1988). Quaternary deposits in the Weihe Basin exceed 1000 m, and the sediment records indicate subsidence rates of about 1mm/yr (Zhang et al., 1998). The central grabens in the SRS, including the Linfin and Taiyuan basins, started to develop in the Neogene, and the north Shanxi grabens started to develop in the late Neogene. The chronology of the Cenozoic sediments found within these basins was mostly determined from vertebrate fossil remnants. Red clays, sands, and gravels are the oldest sediments; up section are lacustrine sediments, and the youngest deposits are eolian loess (Li et al., 1998). In summary, there is a general
consensus that rifting began in the southern parts of the SRS in the Eocene/Oligocene, and significant rifting occurred in the Pliocene to Pleistocene (Li et al., 1998; Zhang et al., 1998; Lei, 2012).

The thicknesses of sediments in these basins suggest in total fault throw, and the ages and stratigraphy of these sediments can help ascertain variations in fault initiation throughout the rift zone. In the northern and southern grabens, the thickest deposits are found in the southeast parts of the respective basin. In the central grabens, deposits are thickest in the middle or northwest parts of the basins (Figure 2.4). With this knowledge of maximum sediment thickness in relation to the location within the basin, estimations can be made in regards to the behavior of graben block movement throughout time.

Volcanism has also played a major role in understanding the timing of events in the SRS. Volcanic activity in the area is thought to be a product of slab rollback of the subducting Pacific plate (Zhang et al., 2003). In the Datong Basin (Figures 2.4 and 2.5), basaltic lavas are found between layers of Pleistocene loess, and the sediments baked from these lavas were dated to confirm the ages of the lavas to the Pleistocene (Li et al., 1998). This indicates that the Pleistocene was a time of intense volcanic eruption in this area.

While the timing of basin formation in the Shanxi Rift System is generally understood, much work is still needed to elucidate the processes involved in the extensional tectonics of North China. The tectonics in North China were originally thought to be solely the result of the extrusion of China (“escape tectonics”) from the collision of India and Eurasia (Tapponnier and Molnar, 1977). However, other possible
Figure 2.4  Map of the Shanxi Rift System with labeled basins, fault information, and structural profiles across three basins. Isopach lines show thickness of Pliocene-Quaternary sediments. Note how the sediment collects in the southeast of the northern and southern basins, while the middle basins, the sediment collects in the northwest of the basin. Location shown in Figure 2.1. Figure modified from Li et al., (1998).
Figure 2.5  Evolution of the Datong Basin as a model for basin development in the Shanxi Rift System.  1-gneiss, 2-limestone, 3-volcanic rock, 4-basalt, 5-residue, 6-reddish clay, 7-loess (Q2), 8-loess (Q3), 9-gravel, 10-clay, 11-fault.  Figure modified from Li et al., (1998.)
influences such as subduction of the Pacific plate and gravitational collapse of the Tibetan Plateau should be considered (Liu et al., 2007).

The timing of initiation of subduction of the Pacific plate on the eastern side of the Eurasian plate is still poorly understood, but based on the analysis of Mesozoic igneous rocks, it is proposed that the eastern side of Eurasia became an active margin sometime before the Jurassic period and that the subducted slab then became stagnant at the mantle transition zone (\(\sim 410-660 \text{ km}\)) (Zhao et al., 2013 and references therein). The subduction of the Pacific plate induced mantle flow underneath east China during the Mesozoic to Cenozoic, which led to widespread magmatism and extension across eastern China, but the effects of the Pacific plate subduction diminished towards the west (Zhang et al., 2003; Zhao et al., 2013, and references therein). This subduction process can explain the extensional forces and volcanism in the early Tertiary, but appear to be inconsistent with current forces and dynamics seen near the SRS today, as suggested by predicted velocities derived from tectonic modeling (Liu et al., 2007).

Over time, the direction of the tectonic stresses acting upon the SRS have changed. Paleostress studies of Zhang et al. (2003) suggest at least two different directions of extension. (Figure 2.6). One set of fault slip data crosscuts rock units that have an age range from prior to the Tertiary to the Miocene, and is oriented NNE-SSW; the second set crosscuts late Quaternary loess and is oriented NNW-SSE. The latter fault-slip data are consistent with the mean extensional directions derived from individual focal solutions of local earthquakes (Zhang et al., 1998, and references therein). Zhang et al., (1998) analysis of 13 earthquakes, ranging from magnitude 3.5 – 5.2, show that the deviatoric horizontal axis for the north Shanxi grabens is approximately 153° (Figure
Figure 2.6  (A): Map showing major faults, earthquake locations, extension directions, and directions of stress calculated from fault slip directions.  (B): Lower-hemisphere stereonets showing fault slip vectors and extension directions. Figure from Zhang et al., (2003).
Zhang concluded that the distinct change of extension directions is a possible result of changes in plate boundary forces, and not due to rotation of tectonic blocks (Zhang et al., 2003).

Numerical geodynamic models have attempted to assess possible forces acting upon the SRS. Using GPS and neotectonic data, Liu et al., (2007) used a three-dimensional finite-element model with rheological and boundary conditions to determine the main tectonic forces acting in China. The subduction zone to the east, the India-Eurasian collision, and the topographic loading forces of the Tibetan Plateau on surrounding areas were examined. Their best-fit models suggest that the forces related to the collision of India and Eurasia are largely influenced by the Tibetan Plateau. The synthetic velocity fields generated by the model for forces related to the topographic loading and gravitational spreading of the Tibetan Plateau were similar to the observed GPS velocity fields. This may indicate that the forces related to this process have a far-reaching impact on the tectonics of eastern Asia. Forces related to subduction in the east had little impact on the predicted deformation in North China, therefore subduction may not currently play a large role in the tectonics of eastern China compared to forces related to subduction in the early Tertiary.

The historical earthquake record of China goes back thousands of years. The earliest recorded documentation of a destructive earthquake in the Shanxi Rift System dates back to 231 AD. Since then, 19 magnitude 6 or larger events have occurred, including one M=8, five M=7-7.9, and thirteen M=6-6.9 earthquakes (Xu et al., 1993a; Li et al., 1998). In 1556, an earthquake in Huaxian killed more than 800,000 people. Compared to other intraplate settings, this area of the world exhibits high amounts of
Figure 2.7  Lower-hemisphere stereonet projection of an analysis of 13 earthquakes in the northern grabens of the Shanxi Rift System. The small black arrows indicate slip vectors on normal faults; the larger black arrows indicate orientation of the tensional axis ($\sigma_3$). Figure from Zhang et al., (1998).
seismicity. However, because large earthquake events are infrequent, the pattern of seismicity is still not well known. At most plate boundaries, earthquakes exhibit some periodicity, even though the spatial variation may seem random (Liu, et al., 2011). However, earthquakes in intraplate settings may exhibit complex behavior, occur sporadically, or have no spatial pattern. For example, earthquakes in the Weihe and Shanxi Riffs exhibit some periodicity over time due to constant slip rates, compared to faults within the North China Plain that exhibit no periodicity because the tectonic forces are loaded onto a complex network of faults (Figure 2.8) (Liu et al., 2011).

III. Study Area

The study area is located on the northern end of the Shanxi Rift System, in the Daixian-Fanshi Basin, which is bounded by the North Wutaishan fault (Figure 2.9). Quaternary sediments are located in the basin, and mountainous areas are comprised mainly of Precambrian igneous and metamorphic units, with occasional Cambrian sedimentary units (Figures 2.10 and 2.11). Specifically, the mountainous areas are comprised of the metamorphic Archean Fuping Hengshan Complex and Proterozoic Wutai Group; other lithologic units in the area include the metamorphic Proterozoic Changcheng System and Hutuo group, undifferentiated Cambrian and Jurassic lithologies, and the Paleogene Fanshi Basalt.

The Wutaishan fault is a normal fault that strikes 60°-70°, dips 45°-80° toward the north, and is about 85 km long (Yu, 2004). There is also evidence of smaller faults splayed off of the main fault. The Wutaishan fault forms the border of the southeastern portion of the Daixian-Fanshi Basin, which is a half-graben basin (Figure 2.12). In the Daixian-Fanshi Basin, the deposition of sediment occurs north of the fault, in the hanging
Figure 2.8  Comparison displaying the differences in seismicity patterns for the a) Weihe and Shanxi Rift and b) comparison of the Weihe-Shanxi rift and North China Plain.  Note in the first comparison, when an increase of events occur in one rift, the amount of events in the other rift decreases.  However, in the North China Plain, there is no pattern of increase or decrease of seismicity when the Shanxi-Weihe system exhibits seismicity.  Figure from Liu et al., (2011).
Figure 2.9  Structural geology map including active faults and seismicity around the Xinding and Daixiain-Fanshi Basin and study area. 1) Normal Fault; 2) Reverse Fault; 3) Strike-slip fault; 4) Inferred fault; 5) Basin border; 6) Oligocene basalt; 7) Quaternary; 8) Pliocene; 9) Cambrian-Ordovician limestone 10) Precambrian metamorphic rock 11) $7 \leq M < 8$ earthquake; 12) $5 \leq M < 7$ earthquake. Star denotes study area. Figure from Xu et al., (1993b).
Figure 2.10  Geologic map and interpreted profile of the study area and surrounding region.  
Figure 2.11 Geologic units found near the study area and surrounding region. Legend on the following page. Figure modified from Ma et al., (1991).
Legend for geologic map and stratigraphic column. Figure modified from Ma et al., (1991).
Figure 2.12  Block model of the formation of some of the half-graben basins in the Shanxi Rift System, specifically, the Daixian-Fanshi Basin. Only one side of each half-graben is actively faulting and relatively uplifting compared to the other side of the basin. Figure modified from Xu et al., (1993a).
wall; mountain uplift occurs south of the fault, in the footwall (Figure 2.13). Fault uplift
rates calculated from vertically displaced fluvial terraces in the area range from 0.6 to
2.00 mm/yr (Yu, 2004; Rui, et al., 2010).

The formation of terraces along the Yangyan River and Wutaishan Fault can be
attributed to the tectonics in the area, and also to regional climatic changes. Tectonic
uplift, climate change, and river incision/aggradation has produced a total of seven
distinct terraces and seven pediment or erosional surfaces that have been regionally
correlated along the Yangyan River based on paleosol and loess deposits using relative
and thermoluminescence dating and utilizing the marine $\delta^{18}$O record (Figures 2.14 and
2.15) (Imbrie et al., 1984; Heslop et al, 2000; Zhang et al., 2007). While the relative
positions of these terraces to each other can be attributed to tectonics, the creation and
stratigraphy of these terraces can be traced back to distinct climate stages, such as glacial
cycles. Loess and soils can be attributed to cold and dry conditions (glacial) or more
warm and wet conditions (interglacial), respectively. The Yangyan River is
approximately 20 km long and flows into the Hutuo River. The deposits along the
Yangyan range from clastic sediments to wind-blown sediments (loess). The pattern of
deposits of the Yangyan terraces also indicate past climate. For Terrace 1 (T1), the well-
sorted binary deposits indicate the river was in an equilibrium state; for T2, the rhythmic,
poorly-sorted deposits indicated a state of aggradation; and for T3-T5, aeolian soils
directly overlie the alluvial gravels, thus indicating a degradational state of the river
(Zhang et al., 2007).
Figure 2.13  Shallow seismic reflection profile of the Wutaishan Fault at Ekou Village. E3-Oligocene basalt; N2 Pliocene; Q1 lower Pleistocene; Q2+3, middle and upper Pleistocene. Seismic velocities are noted for each layer. Figure from Xu et al., (1993b).
Figure 2.14  Generalized stratigraphic sections and relative position of terraces along the Yangyan River. Figure from Zhang et al., (2007).
<table>
<thead>
<tr>
<th>Terrace</th>
<th>Height above valley floor (m)</th>
<th>Alluvial deposit</th>
<th>Thickness of alluvium (m)</th>
<th>Thickness of loess (m)</th>
<th>Oldest loess</th>
<th>Age of terrace (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>T1</td>
<td>4.2</td>
<td>lower section is sand and gravel, subangular to sub-rounded, loose; upper section is silt and fine sand</td>
<td>4.2 (visible)</td>
<td>None</td>
<td>None</td>
<td>0.006</td>
</tr>
<tr>
<td>T2</td>
<td>11—18</td>
<td>sand and gravel, subangular to subrounded, poorly sorted, loose</td>
<td>8—15 (visible)</td>
<td>3.0</td>
<td>L1</td>
<td>0.02</td>
</tr>
<tr>
<td>T3</td>
<td>30</td>
<td>sand and gravel, subangular to subrounded, obvious calcareous cementation</td>
<td>6—8</td>
<td>2.5—8.0</td>
<td>S1</td>
<td>0.13</td>
</tr>
<tr>
<td>T4</td>
<td>96</td>
<td>sand and gravel, subrounded to rounded, multi-rhythmic, well sorted, strong calcareous cementation</td>
<td>14—18</td>
<td>27.9</td>
<td>S5</td>
<td>0.60</td>
</tr>
<tr>
<td>T5</td>
<td>122</td>
<td>sand and gravel, subrounded to rounded, strong calcareous cementation</td>
<td>8—10</td>
<td>45.7</td>
<td>S15</td>
<td>1.20</td>
</tr>
<tr>
<td>T6</td>
<td>250</td>
<td>relic gravels, subrounded</td>
<td>none</td>
<td>3—5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>T7</td>
<td>310</td>
<td>relic gravels, subrounded</td>
<td>none</td>
<td>3—5</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Figure 2.15**  The seven terraces along the Yangyan River. Height above valley floor is the height of the alluvium-loess contact. Table from Zhang et al., (2007).
Chapter 3: Paleostream Profile and Terrace Analyses

While the cumulative effects of tectonic deformation in a region, such as mountains, can easily be perceived, the rates of deformation and tectonic activity in an area may be harder to quantify. Watersheds and streams record the effects of tectonic activity and bridge the time gap between hard rock features and instantaneous methods of calculating deformation. In this chapter, longitudinal stream profiles and stream terraces are examined to qualitatively and quantitatively analyze tectonic uplift and deformation. More specifically, stream profiles and fluvial terraces can provide information about the magnitude, spatial variability, and geometry of tectonic deformation in a region.

With so many variables that can influence streams and their terraces, some assumptions are required to study these features in a particular region. Assumptions made for this study include:

- that there is no significant spatial variation of climate within the region
- there is no fluctuation of base level change throughout time from an outside source that could influence the stream profile or terraces in the study region
- the basic shape of the stream profile has stayed constant throughout time
- rock resistance is relatively uniform (Merritts and Vincent, 1989).

The study area is small enough to be under the influence of the same climate systems and changes throughout time. In regards to base level change, there are two processes that can result in base level change of a river: tectonic uplift or changes in sea level (Merritts et al., 1994). Tectonic uplift in an area increases the relief, which increases the gravitational potential energy for a stream. The stream's response to this uplift and a lowered base level is incision. The relative downward movement of the fault
blocks causes a change in base level for streams in the area, which results in the formation of fluvial terraces and erosional surfaces, depending on mountain uplift/basin subsidence rates and fluvial conditions, such as an aggrading or incising fluvial system (Figure 3.1). It is assumed that the study area is too far inland to be affected by fluctuations in sea level.

I. Methods

a. Stream Profile Construction and Paleostream Profile Reconstruction

Paleostream profiles of the Yangyan River, and other streams, were reconstructed using terrace data gathered from both field measurements and digital elevation models (DEM) utilizing a geographic information system (GIS). In the field, locations of relict terraces were recorded using a handheld GPS unit (Garmin eTrex30), which provides horizontal accuracy of about 5 m or less in ideal conditions. WGS84 was used for the spheroid. Terrace heights above the stream channel were measured by using a Pretel ALTIplus K2 altimeter, accurate to ±1 m, which was calibrated to the same elevation at the same location daily to correct for changes in atmospheric pressure. From the beginning to the end of a work day, the altimeter varied from the initial daily calibration of about ±11 m, which resulted from changes in barometric pressure. To reduce this uncertainty, the terrace elevation and the corresponding stream channel point's elevation were measured as close in time as possible. Terrace height above the stream channel was also measured by using a 2 m height rod and laser rangefinder.

To prepare these terrace data for profile reconstruction analysis, the current stream profile was extracted from a DEM to compare altimeter-based profiles and to create a profile for the entire length of the stream. The DEM used for analysis was from
Figure 3.1 The formation and evolution of fluvial terraces in the footwall of a normal fault. T – Terrace; P – Pediment. Figure from Zhang et al., (2007).
the Advanced Spaceborne Thermal Emission and Reflection Radiometer Global Digital Elevation Model (ASTER GDEM) (http://gdem.ersdac.jspacesystems.or.jp/) dataset, and processing utilized ArcGIS 10. The DEM was conditioned using the Optimized Pit Removal tool, which can be found at: http://tools.crwr.utexas.edu/index.html. This specific tool balances the filling and carving of spurious/artificial pits and peaks found within the DEM (Figure 3.2). An overview of the optimization technique is given by Soille (2004). The importance of conditioning the DEM is to create a hydrologically "correct" surface, in which water will flow downhill and not collect in artificial sinks. Having a hydrologically correct DEM is necessary for proper stream profile creation. HydroSHEDS SRTM (Hydrological data and maps based on the SHuttle Elevation Derivatives at multiple Scales; Shuttle Radar Topography Mission; http://hydrosheds.cr.usgs.gov/index.php) data were also used to compare to the GDEM and satellite imagery, but it was found that GDEM data provided the most accurate stream delineation in relation to streams featured in satellite photographs.

After the DEM was conditioned, the stream channels were delineated within the DEM. The flow direction, flow length and flow accumulation of the DEM were then calculated. With the flow accumulation results, a conditional statement was implemented to create a raster displaying only the pixels of probable stream networks. The raster dataset was converted into polyline segments that could be selected for individual stream analysis, if needed. The polylines or stream raster dataset could then be used as a mask to extract elevation data from a DEM. The elevation data was used on the Y-axis of the stream profiles. X and Y coordinates for the stream points were
Figure 3.2  Three methods of correcting a Digital Elevation Model to make it hydrodynamically correct for use in making stream profiles and other analyses. The pit and "peak" may be artificial remnants of the DEM creation process. The Optimized Pit Removal tool balances the conditioning of a DEM between filling potentially artificial pits and removing peaks. Figure from the Optimized Pit Removal tool tutorial, which can be found at http://tools.crwr.utexas.edu/index.html, with overview given by Soille (2004).
manipulated by using the distance formula to calculate downstream distance from the headwaters for the X-axis of the stream profile.

The longitudinal profile can now be graphed with the elevation and distance from headwater information (Figure 3.3). Significant breaks or curvatures that deviate from the profile can either represent artificial remnants from the DEM extraction, or the features can represent actual changes in topography. These breaks, if natural, may allude to the location of knick points due to tectonic activity or change in lithology.

To construct a paleostream profile, this study utilized stream terraces which possess both age and elevation data. Stream terrace elevations, and, if possible, heights above stream were recorded. Proximity to other terraces and terrace features were also recorded, such as alluvium/loess contacts, alluvium thicknesses, etc. If the height above the stream valley and age of the terrace is known (Figure 2.15), as well as the position along the stream, then an approximate paleostream profile can be created by "connecting the dots" of terraces of similar age. By plotting these terraces in approximate relation to the current stream, then the paleostream profile and modern stream profile can be compared. For the Yangyan River and its terraces, the terraces were added to the modern stream profile graph by adding the altimeter-measured height above current stream to the profile generated from the DEM (Figure 3.4).

b. Profile Adjustment and Error Calculation

Since the elevation data of the DEM-generated stream profiles and the altimeter-generated profiles/terraces did not initially use the same datum, both profiles did not match up when plotted together, so a method was needed to correct them to the same datum. For this thesis, the altimeter-generated profiles were corrected to the DEM-based
Figure 3.3  Stream profile of the Yangyan River generated from GDEM data. The plotted profile has a vertical exaggeration of 8X. The mountain front is marked, indicating the mountain/basin transition and location of the Wutaishan fault.
Figure 3.4  Stream profile of the Yangyan River generated from GDEM data. Also plotted are stream terrace points measured out in the field. The plotted profile has a vertical exaggeration of 8X. The mountain front is marked, indicating the mountain/basin transition and location of the Wutaishan fault.
datum. This was accomplished by adding a constant value in order to minimize the root mean square error (RMSE) between the DEM profile and the adjusted altimeter profile. The RMSE is a value that describes the relationship between two sets of numbers; when the RMSE value approaches zero, the compared datasets are more similar. The basic formula used to calculate the RMSE for a particular stream was:

\[
RMSE = \sqrt{\frac{1}{n} \sum_{i=1}^{n} (\hat{Z}_i - Z_i)^2}
\]  

(3.1)

where, in the context of this thesis, \(n\) is the number of elevation points used in the RMSE calculation; \(\hat{Z}\) is the elevation from the SRTM DEM dataset; and \(Z\) is the adjusted altimeter elevation. It is this \(Z\) value that has to be manipulated to determine the optimum RMSE.

The first step in calculating the error value to adjust the altimeter/terrace datum was to average 3-4 altimeter data points that were matched to the closest DEM data point. Preferably, these would be two points behind and two points in front of the DEM data point. By using this moving window of two in front and two behind, the overlapping windows makes the averaging of the altimeter points and eventual DEM profile comparison more robust. For all of the averaged altimeter points, there should be a corresponding DEM-generated profile point. Then, the sum of the averaged altimeter points are subtracted from the sum of the elevations of the matched DEM profile points. This value is then squared, and then divided by the total number of corresponding/matched DEM profile points. Then, the square root of this value is taken. This value is the RMSE for the unadjusted average altimeter data points.
In order to get the RMSE as close to zero as possible, the averaged altimeter elevation data must be adjusted as a whole. When the optimal RMSE value is reached, the adjustment value used to attain that RMSE was applied to the altimeter stream profile data. The optimization of the RMSE and subsequent elevation adjustment produces a profile, originally based on altimeter data, that statistically fits the DEM-derived stream profile.

If the terrace height above the stream was measured using the height rod or laser rangefinder, then the final terrace elevation value was based on the adjusted stream height plus the measured height above the stream. If the terrace's height above the stream wasn't directly measured (i.e., only used the altimeter), then the terrace elevation was adjusted solely based on the adjustment value of the stream.

c. Cartosat DEM Surface Profile Methods

To analyze and interpret subtle variations in the topography in the study area, the study used a 5 meter pixel DEM generated from stereoscopic Cartosat imagery. The vertical resolution that is comparable to the GDEM. Four topographic profiles were extracted using ERMapper. The four profiles consisted of a profile east of Stream 1 (E1); west of Stream 2 (W2); between Stream 2 and Stream 3 (2/3); and East of Stream 3 (E3) (Figure 3.5).
Figure 3.5 Overview of the study area and the streams that were examined. The Altimeter/GPS profiles, labeled 1-4, were streams examined west of the Yangyan River, and are tributaries of the Hutuo River. The surface profiles generated were east of Stream 1 (E1), west of Stream 2 (W2), between Streams 2 and 3 (2/3), and East of Stream 3 (E3). Terrace points collected are denoted with a circle; interpreted terrace transitions are denoted with a triangle.
II. Results

a. Yangyan River

The stream profile of the Yangyan River from the GIS is approximately 37 km long, and was used in conjunction with 42 terrace spot heights for paleoprofile interpretations (Figure 3.6 and Plate 1). All data points were collected in the footwall portion of the Wutaishan fault. The terrace stratigraphy from Zhang et al., (2007) (Figure 2.15) was used for identifying terraces along the Yangyan River. There are remnants of all seven terraces in the Yangyan watershed, but not all are prevalent throughout the whole length of the river.

North of the Wutaishan fault, in the hanging wall block, there were no terrace data points measured because they were buried by basin fill. However, in the footwall block, surfaces of Terraces T1-T7 were measured. There was one terrace point for T6; one terrace point for T7; two terrace points for T4; and two terrace points for T5. There were several terrace points that were interpreted as T2 and T3. The interpreted paleoprofile for T2 extended approximately 12km upstream from the mountain front. The interpreted paleoprofile for T3 extended approximately 13 km upstream from the mountain front.

Although there were no terrace data points collected in the hanging wall to determine subsidence rates, there is a possible upstream convergence of the T2 and T3 paleoprofiles that may be indicative of tectonic activity in the area (Plate 1). This convergence may be a result of fault block tilting.

b. Stream 1

Stream 1 was the westernmost stream analyzed in the study area. The profile extracted from the DEM was approximately 6.3 km long starting from the headwaters of
Figure 3.6  Longitudinal profile and terraces of the Yangyan River. The mountain front/basin transition and location of stream order transitions are marked. Dashed lines connecting terrace data points are interpreted paleoprofiles for T2 and T3 terraces. The plotted profile has a vertical exaggeration of 8X. The Yangyan River continues approximately 11km past the mountain front and drains into the Hutuo River. See Plate 1 for full profile.
Stream 1, and the stream profile created from an altimeter/GPS survey was approximately 2.3 km long. The altimeter/GPS profile had an elevation change of 125 m. Four data points were collected for terrace analysis (Figure 3.7 and Plate 2). The stream profile based on the altimeter data had to be adjusted -18 m to best fit the DEM stream profile, with an RMSE value of 8.13 meters.

North of the Wutaishan fault, in the hanging wall, there was one each of a T1, T2, and T3 surface. South of the Wutaishan fault, in the footwall, there was one data point collected, which may represent a remnant of a T3 surface.

c. Stream 2

Stream 2 has a profile that is approximately 4.6 km in length. Field measurements spanned 1.9 km in length and included 22 terrace points (Figure 3.8 and Plate 3). The altimeter/GPS profile had an elevation change of 153 m. The stream profile based on the altimeter data had to be adjusted -17 m to best fit the DEM stream profile, with an RMSE value of 7.28 meters.

North of the fault, in the hanging wall, there was one series of correlated T1 surfaces, and four concordant T2 surfaces. South of the fault, in the footwall, there was one series of correlating T2 surfaces. Also in the footwall was an interpreted T1 surface and possible T4 surface. It is interpreted that there is an approximately 11 m offset between the T2 surfaces on either side of the fault. The age of T2 is reported as ~20 ka ±0.9 ka (Zhang et al., 2007); this suggests the rate of fault throw for this offset would be about 0.53 – 0.58 mm/yr.
Figure 3.7   Longitudinal profile and terraces of Stream 1, the westernmost stream examined in this thesis. The profiles consists of a profile from GDEM data and the adjusted altimeter profile. The terraces have also been adjusted to the GDEM datum. The mountain front/basin transition and location of a fault splay is marked. Dashed lines connecting terrace data points are interpreted paleoprofiles for T2 and T3 terraces. The plotted profile has no vertical exaggeration. See Plate 2 for full profile.
Figure 3.8  Longitudinal profile and terraces of Stream 2. The profiles consists of a profile from GDEM data and the adjusted altimeter profile. The terraces have also been adjusted to the GDEM datum. The mountain front/basin transition and location of a fault splay is marked. Dashed lines connecting terrace data points are interpreted paleoprofiles for T2 and T4 terraces. The plotted profile has no vertical exaggeration. See Plate 3 for full profile.
**d. Stream 3**

Stream 3 consists is approximately 2.5 km in length. Field measurements were made over one kilometer in length and included 18 terrace points (Figure 3.9 and Plate 4). The altimeter/GPS profile had an elevation change of 88 m. The stream profile based on the altimeter data had to be adjusted -4 m to best fit the DEM stream profile, with an RMSE value of 14.00 meters.

North of the fault, in the hanging wall, there is one interpreted series of correlated T2 surfaces, a T3 surface, and a correlated series of T1 surfaces that extend from the hanging wall and into the footwall. South of the fault, in the footwall, there are T1 and T2 surfaces that can be correlated past the mountain front and into the hanging wall. There is an interpreted 6 m offset between the T2 surfaces that are on either side of the fault. Using the estimated T2 age of ~20 ka ±0.9 ka (Zhang et al., 2007), the throw rate for this offset would be about 0.29 – 0.31 mm/yr.

**e. Stream 4**

Stream 4 is approximately 4.3 km in length. Field measurements spanned 1.6 km in length and included 52 terrace points (Figure 3.10 and Plate 5). The altimeter/GPS profile had an elevation change of 160 m. The stream profile based on the altimeter data had to be adjusted +2 m to best fit the DEM stream profile, with an RMSE value of 9.63 meters.

In the footwall of the fault, there was one series each of T1, T2, and T3 surfaces that could be correlated. There were no terrace data points collected in the hanging wall of the fault (north of the mountain front).
Figure 3.9  Longitudinal profile and terraces of Stream 3. The profiles consists of a profile from GDEM data and the adjusted altimeter profile. The terraces have also been adjusted to the GDEM datum. The mountain front/basin transition and location of a fault splay is marked. Dashed lines connecting terrace data points are interpreted paleoprofiles for T2 and T3 terraces. The plotted profile has no vertical exaggeration. See Plate 4 for full profile.
Figure 3.10  Longitudinal profile and terraces of Stream 4. The profiles consists of a profile from GDEM data and the adjusted altimeter profile. The terraces have also been adjusted to the GDEM datum. The mountain front/basin transition is marked. Dashed and solid lines connecting terrace data points are interpreted paleoprofiles for T1, T2, and T3 terraces. The plotted profile has no vertical exaggeration. See Plate 5 for full profile.
f. Other DEM Surface Profiles

Surface profiles of the fault splay block were extracted from the 5 m pixel DEM. These profiles were examined to assess subtle changes in topography in the fault splay block west of the Yangyan River. This fault splay block is delineated by the Wutaishan Fault to the south, and a smaller unnamed fault to the north (Figure 3.5). In all four generated profiles, there is indication of an inflection in the surface. This inflection is interpreted to correspond with the riser between the T2 and T3 surfaces because there was no evidence of faulting or folding of sediments viewed in the stream cut near this riser (Figures 3.11 – 3.14). This is most evident on E1 and W2.

g. Fault Throw Rates Calculated from Previous Works

Utilizing terrace height data from Zhang et al., (2007) and sediment thickness data derived from a shallow seismic profile Xu et al., (1993b), fault throw and fault throw/uplift rates were calculated for five terraces (Figure 3.15 and 3.16). The shallow seismic profile in Figure 2.13 was corrected to have no vertical exaggeration, and the total thickness of Quaternary sediments in the hanging wall was averaged. The seismic profile shows the boundary between Neogene and Quaternary material in the hanging wall, so by using the averaged Quaternary thickness, it was calculated that this section of the Wutaishan fault had a sedimentation rate of approximately 0.12 mm/yr for the duration of the Quaternary. The sedimentation rate was used in conjunction with the ages of terraces 1-5 (Figure 2.15) to calculate an estimated burial depth of the sediments that have a reverse stratigraphy of the terraces in the footwall. These burial depths, along with the terrace heights above the river valley floor were summed to calculate the throw of the fault for the five terraces. The total throw and age of each terrace was used to
Figure 3.11 Longitudinal profile of a topographic surface east of Stream 1. A possible inflection in topography indicating a transition from Terrace 2 to Terrace 3 is labeled.
Figure 3.12  Longitudinal profile of a topographic surface west of Stream 2. A possible inflection in topography indicating a transition from Terrace 2 to Terrace 3 is labeled.
Figure 3.13  Longitudinal profile of a topographic surface between Stream 2 and 3. A possible inflection in topography indicating a transition from Terrace 2 to Terrace 3 is labeled.
Figure 3.14  Longitudinal profile of a topographic surface east of Stream 3. A possible inflection in topography indicating a transition from Terrace 2 to Terrace 3 is labeled.
Figure 3.15  Fault throw and fault throw/uplift rates calculated from terrace height and age data from Heslop et al., (2000) and Zhang et al., (2007) and shallow seismic data from Xu et al., (1993b).

<table>
<thead>
<tr>
<th>Terrace</th>
<th>Height above valley floor (m)</th>
<th>Thickness of alluvium (m)</th>
<th>Thickness of loess (m)</th>
<th>Age of terrace (ka)</th>
<th>Potential Age Error (ka)</th>
<th>Estimated burial depth in hanging wall (m)</th>
<th>Estimated Total Throw (m)</th>
<th>Fault throw/uplift rate (mm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>T1</td>
<td>4.2</td>
<td>4.2</td>
<td>none</td>
<td>6</td>
<td>± 0.51</td>
<td>0.7 - 0.8</td>
<td>4.8 - 4.9</td>
<td>0.75 - 0.91</td>
</tr>
<tr>
<td>T2</td>
<td>11.18</td>
<td>8.15</td>
<td>3.0</td>
<td>20</td>
<td>± 0.9</td>
<td>2.2 - 2.5</td>
<td>13.4 - 20.4</td>
<td>0.64 - 1.07</td>
</tr>
<tr>
<td>T3</td>
<td>30</td>
<td>6.8</td>
<td>2.5 - 8.0</td>
<td>130</td>
<td>± 11.42</td>
<td>14.2 - 17.0</td>
<td>14.2 - 17.0</td>
<td>0.31 - 0.41</td>
</tr>
<tr>
<td>T4</td>
<td>96</td>
<td>14.18</td>
<td>27.9</td>
<td>625</td>
<td>± 5</td>
<td>74.4 - 75.6</td>
<td>74.4 - 75.6</td>
<td>0.27 - 0.28</td>
</tr>
<tr>
<td>T5</td>
<td>122</td>
<td>8.10</td>
<td>45.7</td>
<td>1,245</td>
<td>± 5</td>
<td>149 - 150</td>
<td>270.8 - 272.0</td>
<td>0.22</td>
</tr>
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<td>T6</td>
<td>250</td>
<td>none</td>
<td>3-5</td>
<td>Not Known</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>T7</td>
<td>310</td>
<td>none</td>
<td>3-5</td>
<td>Not Known</td>
<td></td>
<td></td>
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<td></td>
</tr>
</tbody>
</table>
Figure 3.16  Fault throw and fault throw/uplift rates calculated from terrace height and age data from Heslop et al., (2000) and Zhang et al., (2007) and shallow seismic data from Xu et al., (1993b). The vertical black bars indicate the range of the fault throw/uplift rate.
calculate the throw/uplift rates throughout the past 1.6 million years. It was observed that, in general, the fault throw/uplift rate was relatively constant at 0.22-0.41 mm/yr from approximately 1.2 – 0.13 Ma. Then, sometime between 0.13 Ma and the present, the rate appears to have greatly increased to a range of 0.64-1.07 mm/yr.

**III. Stream Profile and Terraces Discussion**

In all streams studied, the paleostream profiles are generally parallel to the modern stream profile in the uplifted, footwall block. However, when comparing the paleostream profiles from different age surfaces along the Yangyuan River, some upstream convergence is apparent. This may suggest the fault block south of the Wutaishan fault tilts towards the southeast.

Because rates of fault throw were calculated from terrace measurements in the fault splay block, the subsidence rates cannot be compared directly to the main Wutaishan fault. This is because subsidence rates on the main Wutaishan fault might be divided out between the Wutaishan fault to the south and the smaller fault splay to the north.

Compared to the analysis of Rui et al., (2010), in which the vertical slip rates for the Wutaishan fault west of the Yangyan River were calculated to be ~1.55 – 2.00 mm/yr, the vertical slip rates calculated in this thesis were three to five times less at ~0.3 – 0.55 mm/yr.

The difference in vertical slips rates can be attributed to at least three possible scenarios: 1) the terraces were misinterpreted out in the field; 2) terrace elevations were not accurate due to differences in altimeter, DEM, and real-world elevations; or 3) the smaller fault splay block has a much smaller fault slip/subsidence rate compared to the
main basin/Wutaishan fault. The third scenario is very plausible because the wide T2/T3 surface on top of the fault splay block is still preserved; if vertical fault slip rates were high, and similar to results found in Rui et al., (2010), then this surface may have been buried in the hanging wall of the Wutaishan fault.
Chapter 4: GIS and Morphometric Analyses

Morphometric analyses study landscape shape using numerical approaches. Qualities that describe the landscape such as elevation, size, slope, and their derivatives can be used to quantitatively compare different landscapes to each other. Geomorphic indices are products of morphometric analyses, which may help assess of tectonic activity in an area (Keller and Pinter, 1996). The advantage of performing morphometric analyses, especially with a GIS and DEMs, is that large areas can be studied and compared relatively easily. Scales such as a whole watershed, or hundreds of watersheds, requires techniques that can be scaled up and applied to the entire region with the help of a GIS. Two geomorphic indices that were used in this study are the Stream Length-Gradient Index (SL index) and the hypsometric integral.

I. Stream Length-Gradient Index

As the stream gradient decreases with distance downstream, the power of the stream increases; the SL index reflects stream power, and can be used to compare different sections of the same stream (Hack, 1973). This index can reveal deviations from a stream's idealized equilibrium condition. Since the SL index is related to the slope of the stream, factors that can influence the stream channel over time can also affect the SL index. Factors such as lithologic changes and tectonic anomalies can possibly be detected using the SL index. However, lithologic changes may only influence the SL index at smaller scales, while regional scale tectonic features such as thrust faults have the most influence of the SL index at larger scales (Seeber and Gornitz, 1983; Keller and Pinter, 1996; Chen et al., 2003). For example, two streams in different regions with the
same SL index value does not necessarily mean they are under the influence of the same amount of tectonic forces or have the same lithology.

**a. Stream Length-Gradient Index Methods**

To calculate the SL index for a certain section of a single stream, the formula used by Hack (1973) was used:

\[
SL = \frac{\Delta H \times L}{\Delta L}
\]  

(4.1)

where \(\Delta H\) is the difference in elevation between the ends of the studied segment, \(\Delta L\) is the length of the studied segment, and \(L\) is the length of the stream segment measured from the drainage divide to the midpoint of the studied stream segment (Figure 4.1). Since \(\Delta H/\Delta L\) is the gradient of the stream segment, caution must be used when calculating the SL index near the headwaters of the stream; for most normal streams, the gradient increases greatly near the headwaters of a profile. The SL index may become too high for useful analysis because the headwaters are the least capable of re-grading (Hack, 1973).

In this study, the SL index of a segment of a stream was compared to the idealized SL index of the entire stream length (Seeber and Gornitz, 1983). The above equation could be used to calculate the SL index for the entire length, however, the following equation was utilized for ease of use in the analysis of the SL index for the Yangyan River Case Study and GIS analyses (Hack, 1973; McKeown et al., 1988):

\[
SL = \frac{\Delta H}{\Delta \ln L}
\]  

(4.2)
Figure 4.1 Calculation of the Stream Length-Gradient Index. Variables defined in the text. Figure from Hack (1973).
b. Stream Length-Gradient Index Case Study: the Yangyan River

The normalized SL index for the Yangyan River was calculated using elevation data from the DEM and equation 4.2. To calculate the SL index for the entire Yangyan River, the difference in elevation between the head and the mouth of the stream was divided by the natural logarithm of the Yangyan's length. The distance and elevation difference from the mouth of the Yangyan to approximately 350 m downstream was not considered in this calculation. This omission is because of the potential for error in the SL index calculation due of the extreme gradients near the head (Hack, 1973). The result was an SL index value of 291 for the entire length of the Yangyan.

To calculate the SL index for defined segments of the Yangyan, and to filter out noisy data, a moving window of an approximate length was utilized. For this study, separate moving windows of approximately 500 m and 2000 m were used. In these defined windows, the difference in elevation was divided by the difference in the starting and end point of the window. Since the distances between the pixel points are not divisible by 500 or 2000, the moving window size utilized was approximately 500 m and 2000 m, respectively. The local SL indices generated were then divided by the total SL index for the Yangyan to normalize the data. Then, the logarithm of results was calculated to indicate SL index values greater than the total SL index (positive), values less than the total SL index (negative), or values equal to the total SL index (zero). The Yangyan's profile and the normalized SL index values for the 500 m and 2000 m window were then graphed (Figure 4.2).

It was observed that in the graphed results, for the most part, both results of the 500 m and 2000 m window have the same peaks and valleys. For about the first 6 km,
Figure 4.2 Results of an SL index analysis on the Yangyan River. There were two resolutions of the SL index analysis that consisted of a 500m and 2km traveling window. The mountain front is marked by a circle. There is no vertical exaggeration.
the normalized SL index increased for both windows, and then relatively leveled off. The increase in the normalized SL index for the first 6 km may be due to the fact that the headwaters are least capable of re-grading and reaching equilibrium. Starting at about 16 km, the SL index values start to gradually increase up until about 26 km downstream where the SL index values for both the 500 m and 2000 m window are consistently level and contain the highest average SL index values for a 2000 m window. It is from about 26 km to 30 km downstream, near the mountain front, that values from both windows' values do not vary by much.

**c. Stream Length-Gradient Index GIS Methods**

To calculate the local segment's SL index value in a GIS, the result of a slope analysis was multiplied by the flow length upstream using a raster calculator (drainage divide = zero flow length, end of stream = maximum flow length). This study used stream length segments of 30 m (the size of one pixel in the GIS data) for the SL index calculation. For the amount of data generated, using the 30 m cell size was used for its simplicity in calculating the local SL index with a GIS.

The first step in calculating the SL index for the entire reach of a stream using a GIS was to assign a unique identification number to every stream segment of every Strahler stream order. The resulting raster datasets contain stream orders with all of its attached lower-ordered streams. In this study, there were seven stream orders and their respective watersheds generated and examined. Then, a Zonal Statistics command was performed to determine the minimum and maximum downflow length of each watershed. These points should be the headwaters and mouth of a stream/watershed. The elevations of these points were then extracted from the DEM. Using equation 4.2, all four variables
were then utilized to calculate the SL index for the entirety of each stream reach in all seven stream orders.

The local SL index raster was then divided by the SL index for the entire stream reach for each Strahler stream order (orders 1-7), to normalize the SL index values to a value of one (1). This way, the local SL index values for each segment – 30 meter-long pixels in the GIS – can be compared to the SL index of the entire stream. For example, if the local segment SL index is higher than the entire stream's SL index, then the normalized value will be higher than one. A value higher or lower than one may indicate an anomaly for that particular section of stream; the anomaly may be due to, but not limited to, DEM inconsistencies, tectonics, or change in lithology.

The normalized SL index data was then filtered to average the data and to remove potentially erroneous values. A moving 3x3 pixel kernel was used in a one/null stream raster (stream = one, no stream = null) to find the median normalized SL index value in the kernel window for streams equal to or greater than two pixels in length. This process filters out spurious streams that are only one pixel in length. Next, an Inverse Distance Weighting (IDW) function was used in the GIS to interpolate areas where there is not an SL index value calculated in the mountainous areas. The SL index was only calculated in the mountainous areas, as opposed to the basin, because the subtle variations in the topography of the basin is not resolved sufficiently enough in the DEM for SL index analysis.
d. Stream Length-Gradient Index Results from GIS Analysis

The results of the SL index analysis were divided into second, third, and fourth-order watersheds (Figures 4.3 – 4.5). All of the SL index analyses are in reference to the southern mountainous component of the Daixian-Fanshi Basin. The second-order SL indices show no particular pattern of high or low SL index concentrations throughout the region, except for a small patch of high SL index watersheds in the southwestern portion of the region. Streams entering the basin have relative SL index values from low to high, and display no particular pattern or concentration of value across the mountain front/Wutaishan fault. Third-order watersheds have the highest concentrations of high-valued SL index values in the central to western half of the region. Third-order streams entering the basin have the highest values in the western and central portions of the region, and lower values to the east. Fourth-order watersheds also have the highest concentration of high-valued SL index values to the central and western portions, with streams entering the basin having moderate to high SL indices in the central and western portions of the studied region.

There are possible consistencies of high SL index values between all three orders of watersheds. The third and fourth-order watersheds display the highest concentrations of high-value SL indices in the central and western portions of the mountainous areas south of the Daixian-Fanshi basin/Wutaishan fault. In the southwestern area of the region, high SL gradient index values persist through all three stream orders.
Figure 4.3  Results of an SL index analysis on all order 2 streams in mountainous areas for the Hutuo River basin. The SL index for a stream segment has been normalized to the SL index for the entire length of order 2 streams. The highest and lowest values of the normalized SL index value indicate the greatest deviation from the SL index value for the entire length of a stream.
Figure 4.4  Results of an SL index analysis on all order 3 streams in mountainous areas for the Hutuo River basin. The SL index for a stream segment has been normalized to the SL index for the entire length of order 3 streams. The highest and lowest values of the normalized SL index value indicate the greatest deviation from the SL index value for the entire length of a stream.
Figure 4.5  Results of an SL index analysis on all order 4 streams in mountainous areas for the Hutuo River basin. The SL index for a stream segment has been normalized to the SL index for the entire length of order 4 streams. The highest and lowest values of the normalized SL index value indicate the greatest deviation from the SL index value for the entire length of a stream.
II. Hypsometric Integral Analysis

Hypsometric analysis of a watershed describes degree of dissection and drainage development (i.e., the “maturity”) of a fluvial landscape. A hypsometric curve is a cumulative histogram of elevation as a function of area, and the hypsometric integral is the area under the curve of this comparison (Figure 4.6). Convex upward hypsometric curves and higher integral values indicate a more youthful landscape, while a concave upward curve and lower integral values may indicate relatively older or more mature landscapes. The use of hypsometry to describe landscapes was first proposed by Strahler (1952), and Figure 4.7 illustrates an overview on the evolutionary cycles of a landscape and its hypsometric curve. Calculating the hypsometric integrals for watersheds in an area allows the comparison of multiple watersheds, regardless of basin scale. However, the scale-independence of watersheds' hypsometric integrals only occurs in steady-state topography (Cheng, et al., 2012). Factors that can affect the hypsometric curve/integral include uplift rates, erosional resistance of lithological units, exhumation rates, erosion processes, geologic structures, and climate (Chen et al., 2003; Cheng et al., 2012, and references therein).

a. Hypsometric Integral Methods

To create watersheds based on Strahler stream orders, it is necessary to create separate layers of only order 1 streams, order 2 streams, order 3 streams, etc. A watershed is created by using a pour point at the maximum downstream accumulation value for each stream orders' stream segments. To create a pour point with these conditions, the maximum value of accumulation for each unique stream segment was found. The resulting layer consists of pixels that have a maximum accumulation value
Figure 4.6  A graphical representation of the hypsometric curve for a drainage area. The area under the hypsometric curve is the hypsometric integral. Figure from Keller and Pinter (1996).
Figure 4.7  Two different styles of landscape evolution and their respective hypsometric curves. From Strahler (1952) and Ohmori (1993).
(the furthest downstream of a stream segment). These pixels were then converted into points, which were then used as pour points for creating watersheds.

To calculate the hypsometric integral for each watershed of each stream order, a zonal statistics command was performed on the watershed raster layers and the DEM to find the maximum, minimum, and mean elevation for each watershed. With these three values, the hypsometric integral for each basin can be calculated using Pike and Wilson's (1971) formula, which is mathematically identical to the formula for the basin-relief ratio:

\[
HI = \frac{\text{Mean Elevation} - \text{Min. Elevation}}{\text{Max. Elevation} - \text{Min. Elevation}}
\] (4.3)

The hypsometric integrals for second, third, and fourth-order watersheds were further analyzed in ER Mapper. The use of ER Mapper for hypsometric integral analysis allows the user to modify or enhance the graphical display of the integral values. For example, the extreme minimum and maximum values that may be outliers can be de-emphasized, while subtleties of most hypsometric values can be emphasized and categorized by color.

b. Hypsometric Integral Results from GIS Analysis

Results of the hypsometric analysis were divided into second, third, and fourth-order watersheds (Figures 4.8 – 4.11). For this study, only the southern mountainous watersheds of the Daixian-Fanshi basin were analyzed. It was observed that in the second-order watersheds, the highest integral values in the region are concentrated in the central portions of the uplift. Second-order watersheds that are entering the Daixian-Fanshi basin along the Wutaishan Fault have the highest integral values located in the central portion of the rift segment. In third-order watersheds, the highest integral values
Figure 4.8  Hypsometric integral analysis of stream order 2 watersheds.
Figure 4.9  Hypsometric integral analysis of stream order 3 watersheds.
Figure 4.10  Hypsometric integral analysis of stream order 4 watersheds.
Figure 4.11  Hypsometric integral analysis of all stream orders 2 through 4 watersheds.
in the region are found mainly in the central to southwestern portions of the uplift. Third-order watersheds that are entering the Daixian-Fanshi basin along the Wutaishan fault have the highest integral values located in the central portion of the rift segment. In fourth-order watersheds, the highest integral values in the region are located in the southwest portion of the uplift. Fourth-order watersheds that are entering the Daixian-Fanshi basin along the Wutaishan fault have the highest integral values mainly located in the central portion of the rift.

The consistency between all three watersheds is that the highest integral value concentrations are located near the bend in the fault. Along the Wutaishan fault/Daixian-Fanshi basin boundary, there is a concentration of moderately-valued hypsometric integrals in the central portion of the rift.

**III. Morphometry Discussion**

After analysis of the Daixian-Fanshi basin using SL indices and hypsometric integrals, it appears that the highest morphometry values are found near the bend of the fault/basin, towards the west. While there are no significant changes in lithology in this area (Figure 2.10) to explain this concentration of high SL indices and hypsometric integrals, the fault bend may have a significant influence on morphometric indices. If there is a bend in a fault, it is more efficient for the fault to accommodate movement onto a third fault (King and Nábělek, 1985). The results of the SL index analysis for the Yangyan River indicate that faulting has a major influence on the SL index; the SL index was consistently high at the mountain front/basin transition. The SL index value, along with the hypsometric integral, cannot absolutely indicate the tectonic activity or differences in lithology in an area, and must be used in conjunction with other techniques.
In conclusion, faulting accounts for large amounts of deformation in this area, and this influence can be seen by using morphometric analyses to study the land.
Chapter 5: Structural and Kinematic Analyses

I. Fault Kinematic Analysis

Within the North Wutaishan fault zone, fault-slip data (consisting of strike, dip, rake, and sense of slip) were collected along slickensides of mesoscopic faults. The slickenlines measured were found in a fault zone that separates basement rock in the footwall from Quaternary sediments and gravels in the hanging wall (Figure 5.1). Results of the kinematic analysis were later compared to previous work by Zhang et al., (2003). The goal of the fault kinematic analysis was to determine the orientation of horizontal extension at a local scale, and how it relates to regional scale kinematics.

Seventeen measurements of fault-slip data were taken with a Brunton transit; the kinematic analysis was conducted as implemented using FaultKin 7 (Marret and Allmendinger, 1990; Allmendinger et al., 2012). Results are represented as hanging wall slip directions and a pseudo-fault plane solution from a linked Bingham analysis (Figure 5.2). The linked Bingham analysis is a method of determining the average orientations for a distribution of mutually orthogonal three-dimensional directional data, such as kinematic axes (e.g., the P, T, and B axes of an earthquake focal mechanism). The linked Bingham analysis can also be called an unweighted geometric moment tensor summation (Marret and Allmendinger, 1990). The kinematic axes calculated from the faults and slickenlines indicate the maximum extensional axis (axis 1) is towards the NW-SE direction, with the maximum shortening axis (axis 3) in a near-vertical orientation. The direction of the maximum extensional axis is approximately orthogonal to the normal-faulted basin. Additionally, the eigenvalue of the intermediate axis indicates minimal extension or contraction in the NE-SW direction, i.e., there is minimal shearing at a local
Figure 5.1
Location of the fault gouge sample area, denoted by a star (above). Photograph of sample area with delimited fault gouge zone (right). The leftmost portion of the photograph is of bedrock, while the rightmost portion is of silt and alluvium. Note fieldbook for scale.
Figure 5.2   A linked Bingham analysis of the kinematic indicators from a fault gouge zone. The numbers located in the stereonet represent the three kinematic axes. Axis 1-Maximum Extension; 2-Intermediate; 3-Maximum Contraction. The black arrows represent hanging wall slip directions.
scale. All of the slickenlines measured indicated normal faulting, and the geometric moment tensor summation reflects this; the most plausible fault from the pseudo fault plane solution is a normal fault striking 210 degrees. The results of this analysis were then compared to data found in Zhang et al., (1998, 2003).

II. Results Compared to Previous Works

For more complete spatial coverage of Quaternary kinematics, four datasets from Zhang et al., (2003) were also kinematically analyzed. These data come from the Xizhoushan, Wutaishan, Henshan, and Liulingshan faults (Figure 5.3). The re-analysis of the Zhang et al., (2003) kinematic data compare similarly to the results of this study: the direction of extension derived from Quaternary fault kinematic data is approximately NW-SE. The variation of the extension direction from this study compared to the extension directions found in Zhang et al., (2003) may be explained by the fact that this study's sample site was located near the intersection of the main Wutaishan fault and a smaller fault splay. The area near the intersection of two faults has the possibility of creating a complex network of smaller faults of differing orientations. Another possible explanation for the variation in fault/extension orientations is the fact that the study area is located near a portion of a fault that is changing direction. At the sample collection location, the mountain-bounding fault's strike changes from approximately NW-SE to NE-SW. This change in fault direction may add to the complexity of fault orientations in a localized area. The results of the kinematic analysis indicates that the surrounding area is undergoing extension, and the generation of the fault plane solution helps us visualize that. However, at a regional scale, the sense of deformation is different compared to the local scale.
Figure 5.3 Fault plane solution generated from a fault kinematic analysis of this study compared to the results of Zhang et al.'s (2003) fault slip analysis of faults in the northern Shanxi Rift System. HS1 – Henshan frontal fault; WT1 – Wutaishan frontal fault; XZS1 – Xizhoushan frontal fault.
For a regional scale analysis of present-day kinematics, this study also calculates the seismic strain from earthquake data. The North Shanxi graben earthquake dataset comes from Zhang et al., (1998) and consists of 13 earthquakes from an approximately 65,000 km$^2$ area (Figure 5.4). The data was re-analyzed using Stereonet 8 and Faultkin 7 to produce a seismic moment tensor summation, which provides information about the orientation of the principal strain axes (Kostrov, 1974; Marrett and Allmendinger, 1990). Parameters for the seismic moment tensor calculation include 30 GPa for the shear modulus of Earth's crust, and an average earthquake depth of 15 km (which makes region volume 975,000 km$^3$). From the seismic moment tensor summation, it is revealed that the principal axis 1 (T-axis) direction is at 154.8 with a strain value of 3.11 nanostrain. The principal axis 3 (P-axis) direction is at 63.1 with a strain value of -3.03 nanostrain. Utilizing the span of 16.5 years of collected earthquake data, the strain rates calculated for both axis 1 and 3 are 0.19 and -0.18 nanostrain/year, respectively. The generated average focal mechanism for the dataset reveals that the orientation of the P and T-axis is similar to a strike-slip or "wrench" fault, and that there is minimal extension or contraction in the vertical direction (Figure 5.5). However, the earthquakes analyzed originated from different faults, including normal faults, so the generated "strike-slip" focal mechanism represents an averaged dataset. In summary, the data from a localized set of fault kinematic indicators suggest extension as the mechanism for deformation, while a regional scale analysis suggests horizontal shear or "wrenching" as the main type of accommodation in the northern SRS.
Figure 5.4  Location of earthquakes from Zhang et al.'s (1998) analysis used to determine the regional sense of deformation for this study.
Figure 5.5  Fault plane solution generated from a reanalysis of 13 earthquakes, using a moment tensor summation. First analyzed by Zhang et al., (1998). The three numbers in the stereonet represent the three principal strain axes, with 1 and 3 representing the approximate T and P (tension and compressional) axis, respectively.
Chapter 6: Final Discussion and Conclusions

This study used data from multiple scales to investigate the influence of tectonics on features ranging in size from meso- to regional scale. The data utilized include fault kinematic indicators, stream terraces, and regional/basin-wide morphometry data. With the analyzed data, the following questions were attempted to be answered: 1) How does the modern profile of a stream compare to its paleostream profiles?; 2) Can faulting rates or a regional uplift rate be inferred from terrace measurements?; 3) How do fault rates vary along the strike of the fault?; 4) What does the regional morphometry suggest for spatial patterns of tectonism?; and 5) How do the fault kinematics near the study area compare to regional-scale fault and tectonic kinematics?

For the first question, it was discovered that the paleostream profiles created from T1 and T2 terraces generally follow the modern stream profile. If there was any deviation, it is most likely attributed to measurement/adjustment errors. However, there was evidence in the paleostream profiles of the Yangyan River to suggest fault block tilting. The profiles created from the Yangyan's T2 and T3 terraces appear to exhibit convergence upstream. This convergence may suggest fault block tilting to the west. However, this convergence may simply suggest that faulting mostly affects reaches of a stream closest to the fault, and that there is little to no profile readjustment in the headwaters of the stream.

Faulting rates in the area were inferred by utilizing offset terraces found in Streams 2 and 3. There were no terraces measured in the hanging wall of Stream 4 or the Yangyan River to calculate offset, and the results in Stream 1 were inconclusive. It was calculated that there was an 11 m T2 offset in Stream 2, and a 6 m T2 offset in Stream 3;
the subsidence rates associated with these offsets would be 0.53-0.58 and 0.29-0.31 mm/yr, respectively.

Streams 2 and 3 are located in an area delineated by the main Wutaishan fault to the south, and a smaller fault splay to the north. Since there were no faulting rates calculated in this study for the main Wutaishan fault, variations of faulting rates along strike could not be inferred. However, faulting rates from T2 terraces in Streams 2 and 3 can help us understand this small fault block's movement compared to the main fault. The block's faulting rate ranges from 0.29-0.58 mm/yr, while faulting rates of the Wutaishan fault ranges from 1.55-2.00 mm/yr (Rui, et al., 2010). So for the Yangyan River, and the Wutaishan fault immediately west of the river, the faulting rate is much higher compared to this smaller fault that splays off of the main Wutaishan fault. The results of this study suggest that this area needs to be studied more to fully understand the spatial variation of faulting rates in an area and also the role smaller faults play in regards to strain accumulation in the area.

Quaternary faulting rates were also calculated from terrace height and age data to determine if there has been any variation of the faulting rate of the Wutaishan Fault throughout time. Throughout the Quaternary, starting at or before 1.2 Ma, the fault throw/uplift rate was relatively constant at 0.22 mm/yr to 0.41 mm/yr. Then, sometime around or after 0.13 Ma, the rate greatly increased to a range of 0.66-1.07 mm/yr (Figure 3.16). However, if terraces and faults in other areas in the Shanxi Rift System (SRS) are studied, especially in regards to Quaternary fault throw/uplift rates, then the spatial pattern of fault throw/uplift rates can be assessed for the whole SRS.
The influence of fault splays, and at a larger scale, fault bends, in relation to an area's topography was examined using morphometry analyses. By examining the results from Hyspsometric Integral and Stream Length-Gradient Indices (SL index), it is possible that there is a strong correlation between higher morphometric indices and strain accumulation. At the bend of the Wutaishan fault, towards the west, higher values of the Hyspsometric Integral and the SL index were observed. This is because faulting on each side of the bend cannot be accommodated, and must move on a third fault (King and Nábělek, 1985). If there is no third main fault in an area, strain accumulation must occur elsewhere, such as inside the bend on the western end of the Wutaishan fault. These morphometric analyses are good tools for examining the influence of tectonics at a regional scale.

The final question ties in small fault kinematic features analyzed in the field to other kinematic and tectonic analyses from other studies. The results from an analysis of fault kinematic indicators found at an outcrop reveal NW-SE extension; this compares well to the overall structure of the Daixian-Fanshi Basin/Wutaishan fault, which strikes SW-NE. These results were compared to a previous study by Zhang et al., (2003), which was an analysis of faults in the Shanxi Rift System, and found that the extension direction is also approximately NW-SE. However, a reanalysis of 13 earthquakes in North China using a moment tensor summation, originally analyzed by Zhang et al., (1998), reveal that the orientation of the P and T-axis of a fault plane solution is similar to a strike-slip or "wrench" fault. This suggests that this region is undergoing a “wrenching” type of deformation (i.e., a regional, horizontal shear) in response to tectonic forces.
The strain accommodation for the basins of the SRS are locally controlled by normal faults. The evidence for this includes shallow seismic imaging, trench studies, and fault kinematic indicators. However, the strain accommodation in the SRS, as an averaged whole, can be described as having wrenching characteristics. The evidence for this includes a reanalysis of earthquake data in the region. The morphometric analyses performed for this study links the regional and local scales together, and the results show elevated hypsometric and stream length-gradient index values at the fault bend of the Wutaishan fault. These elevated indices found near the fault bend may indicate the influence of this wrenching style of accommodation at the local scale (Figure 6.1).

The Shanxi Rift System is an active, seismogenic intracontinental rift in an intracontinental setting. Other similar areas of the world are the Baikal Rift in Russia, and the Rio Grande Rift. Field techniques and computer analyses assist in inferring the extent and behavior of tectonics in an area. In fact, computer-assisted techniques, such as stream profile generation or morphometric analyses from a DEM, can help focus field studies onto particular areas to test the results and implications of a computer-assisted reconnaissance techniques.

A more detailed morphometry analysis, as well as a more detailed mapping of smaller faults in the Shanxi Rift System, is needed to better understand the influence of tectonics in the area, areas of high strain accumulation, and patterns of seismicity. It is with this understanding that can help with earthquake risk mitigation.
Figure 6.1 The two main types of deformation acting upon the study region: extension and wrenching. The inset illustrates how a region can develop normal faults when the region is undergoing wrenching deformation. Inset from Keller and Pinter, 2002. E – extension; C – contraction.
References


References, continued


References, continued


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Appendix A

Appendix A describes in detail some of the Geographical Information System techniques used in this thesis. ESRI ArcMap 10.0 was used for these analyses. Most of the tools used can be found in the "Spatial Analyst Tools" toolbox in ArcMap.

I. Drainage Basin Delineation

1. Use the Fill tool (Spatial Analyst Tools- Hydrology) or Optimized Pit Removal tool (http://tools.crwr.utexas.edu/index.html) to condition DEM for hydrologic processing.
2. Calculate the Flow Direction and Flow Accumulation.
3. Use a conditional statement in the Raster Calculator (Map Algebra toolbox) to create a stream network raster: con(raster >=100,1,)
4. Calculate stream orders using the Stream Order Tool.
5. Calculate the upstream flow lengths of the streams using the Flow Length Tool. The upstream flow length calculates the length to the top of the drainage divide. So, the most downstream point on a stream will have the highest flow length using the upstream direction of measurement.
6. Use the Raster Calculator to set up a conditional statement to find streams of a certain order.
7. Use the Stream Link Tool to designate each stream of each stream order as a separate entity.
8. Use the Zonal Statistics Tool (Zonal toolbox) to calculate the maximum flow length for each unique stream. The Input Raster will be the stream link layer; the value raster will be the flow length layer; the statistics type will be Maximum. The calculation of the maximum flow length will be done for each stream order.
9. In order to create a watershed, a pour point needs to be created. Use the Raster Calculator to set up a conditional statement to create a layer of points that indicate maximum flow length value for each unique stream segment. The conditional statement can be written as followed:
   Con(order3==3,3)*Con(zonemax3==flowlength,1) The "order3" and "zonemax3" refers to the Streams of Order 3 layer and the Zonal Statistics Max Flow Length for Stream Order 3 layer. The flow length layer can be used for all calculations. Repeat this step for however many stream orders the study area has. This should create a layer of pixels that correspond to the location of the maximum flow length (the most downstream point of a stream).
10. Convert these pixels into points using the Raster to Point tool in the Conversion Tools toolbox.
11. Use these newly created points and a flow direction raster for inputs into the Watershed tool found in the Hydrology toolbox. When you use pour points for each order of streams, the result will be watersheds based on stream order. Thus, the higher stream orders will produce larger watersheds.
Appendix A

II. Hypsometric Integral Calculation

1. To calculate the hypsometric integral of a watershed, you need two pieces of information: the extent of the watershed (this can be a raster layer) and a DEM.
2. Use the Zonal Statistics Tool (Zonal toolbox) to calculate the mean, minimum, and maximum elevation of the DEM for each watershed. The "Input raster or feature zone data" will be your watersheds for each stream order, and the "Input value raster" will be your DEM layer. The three outputs will be used in the calculation of the hypsometric integral.
3. Use the Raster Calculator (Map Algebra toolbox) to calculate the hypsometric integral, using this general formula: \( \frac{\text{shed}_\text{mean}_\text{elev} - \text{shed}_\text{min}_\text{elev}}{\text{shed}_\text{max}_\text{elev} - \text{shed}_\text{min}_\text{elev}} \)
Appendix A

III. Stream Length-Gradient Index Calculation and Filtering

1. To find the local stream length gradient index for a certain reach of a stream, you first need to calculate three variables: the change in elevation over the stream reach, the length of the particular stream reach, and the total length from the highest point of the watershed to the midpoint of the particular stream reach. The change in elevation over the stream reach length is also the gradient of the stream.

2. To calculate the stream length gradient index using a GIS, you need three pieces of information/layers: a stream network raster, slope as percent rise layer, and flow length.

3. Using the Raster Calculator (Map Algebra toolbox), set up a conditional statement so that you multiply the flow length layer by the slope layer (length x gradient = SL). Example: con(streamraster == 1, flowlength)*(slope/100). Dividing the slope percent by 100 will give the gradient/rise vs. run. The result is the Stream Length Gradient for a particular section of a stream. In this study, this happens to be one pixel of the stream raster (30m).

4. To find the Stream Length Gradient Index for an entire reach of stream, you need two pieces of information for a particular stream: the change in elevation, and change in length. The formula for the SL index for the total length of a stream is:

\[
SL = \frac{\Delta H}{\Delta \ln L}
\]

5. In order to separate out single stream orders for the total length SL calculations, each stream order must be separated out from the stream raster. After the streams of a certain order have been separated out, then use the Stream Link tool to give each stream a unique ID.

6. Using a GIS, the total length SL calculation for each stream segment per stream order needs four pieces of information: the DEM, Flow Length from Upstream, Flow Length of Downstream, and Streams of a Certain Order with unique IDs.

7. To find the change in elevation (\(\Delta H\)), the Zonal Statistics tool is used twice to calculate the maximum and minimum downflow length of each stream segment; a conditional statement is used to determine the maximum and minimum elevation if the Flow Length from Downstream is equal to the maximum and minimum downflow length, e.g. con(maxflowlength == flowlengthdownstream, DEM). These two elevation are then subtracted from each other to determine \(\Delta H\).

8. To determine change in Length, Zonal Statistics is used twice on the Flow Length from Upstream to find the minimum and maximum values.

9. Using Raster Calculator, the minimum and maximum elevations are subtracted from each other divided by the natural log of the subtracted maximum and minimum upstream flow length to determine the SL index of a whole stream, e.g. float((maxdownflowlength – mindownflowlength)/(ln(maxupflowlength)-ln(minupflowlength))).

10. The results of step 9 should result in the SL index for streams of a certain order.
Appendix A

III. Stream Length-Gradient Index Calculation and Filtering, continued.

11. The Local SL Index layer is divided by the SL index for streams of a certain order to create a layer of normalized SL values. If a local portion of a stream's SL index value is similar to the SL index value of the entire stream, the result is a lower number; however, if a section of stream's SL index value is higher than the entire stream's SL index value, the normalized value is higher. This may indicate an anomaly for that particular section of stream; the anomaly may be due to data inaccuracy, tectonics, change in lithology, or something else.

12. Filtering the normalized SL index values will average and "smooth out" the values. The first step for filtering is to create a stream raster from the normalized raster e.g. con(isNull(normalized),0,1). This creates a layer of either a stream (1) or no stream (0).

13. Focal Statistics is then used on the stream layer to determine the sum of stream raster pixels surrounding a particular pixel. The No Data pixel points are ignored.

14. Using the normalized (SL Local divided by SL Total Length) SL raster, the Focal Statistics tool is used to find the median value utilizing a 3x3 pixel window. The No Data values are ignored.

15. With the Block Sum and SL Median raster datasets, the Raster Calculator is used to determine the median SL index value for each pixel of a stream network raster, e.g., con(SLblocksum >= 2 & SLsetnull = =1, SLmedian). In this case, only sums greater than or equal to two pixels are examined.

16. The end result is a raster dataset of filtered median SL index values.

17. For graphical purposes, the SL index values will be interpolated for the entire drainage basin. For this study, the interpolation will only include "mountainous" areas.

18. The interpolation of SL index values is performed using the Inverse Distance Weighted (IDW) tool in the Spatial Analyst Toolbox. The input requires points, so a conversion of the median SL index raster to points is needed. For SL index interpolation manipulation, the raster must be converted into a .tif image so ER Mapper can process it.
Plate 1: Yangyan Profile and Terraces

Elevation (m)

Distance From Head (m)

8x V.E.