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Research Article

Stratigraphic modeling of the Western Taiwan foreland basin: sediment flux from a growing mountain range and tectonic implications

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A B S T R A C T

Sediment flux signals in foreland basins preserve a record of tectonics, sea level and climate through erosion and sedimentation. However, longitudinal sediment transport often occurs in foreland basin, thus removing part of the orogenic material flux from foreland basins. Here we use mass balance calculation and stratigraphic simulations of sediment fluxes for the Taiwan orogen to provide a quantified order of magnitude estimate of how much orogenic material may bypass a foreland basin. Our results indicate a significant, potentially more than 50%, mismatch between sediment volume currently preserved in the basin and the amount of material eroded from the orogen since the onset of collision in Taiwan. This supports previous paleogeographic work suggesting that longitudinal sediment transport in the paleo-Taiwan Strait served as a bypass conduit important for the establishment of a steady state orogen. We identify candidate submarine topography in the South China Sea that may preserve Taiwan's missing erosional mass.

1. Introduction

Sediment fluxes within foreland basins exert a primary control on basin architecture involving interactions between tectonics, sea level and climate through erosion and sedimentation (e.g., Castelltort et al., 2015). The orogenic history of many ancient basins has been reconstructed with help

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of the sedimentary record, such as the Alps (Garzanti et al., 2004; Lihou and Allen, 1996), the Pyrenees (Puigdefàbregas et al., 1992; Vergés and Burbank, 1996), or the Himalayas (Garzanti et al., 2005; White et al., 2002), but it is still not well known how much of the orogenic history is eventually preserved and how tectonics, facies and sediment supply to basins are linked (Castelltort et al., 2015).

The western foreland basin in Taiwan is an excellent place to study interactions between tectonics and sediment fluxes because it is very young and still very active. In this basin, the southwestward ongoing oblique collision between the volcanic arc with the continental shelf makes it possible to record the full evolution of the basin deformation (Suppe, 1981) and provides an opportunity to connect tectonics and depositional processes at different stages of the basin's evolution.

The Taiwan orogen is emblematic of the distinct classical evolutionary stages that characterize many ancient foreland basin systems such as the Molasse basin of the Alps (Allen et al., 1991), the Bradanic Trough in the Appennines (Tropeano et al., 2002) or the Solomon Sea in Papua New Guinea (Silver et al., 1991). The western foredeep in Taiwan evolved from an early underfilled stage with relatively deep-water sedimentation to a late balanced-filled stage, where shallow marine environments persist until today (Covey, 1984), despite the enormous amount of sediment supplied to the ocean by the rising Taiwan mountains (Milliman and Kao, 2005; Milliman and Syvitski, 1992).

It is well known that the Taiwanese collision formed a time-transgressive southwestward oriented migration of facies belts (Nagel et al., 2012c) and sediment depocenters (Simoes and Avouac, 2006), similar to other oblique collisions (e.g. Papua New Guinea, Abbott et al., 1994; Silver et al., 1991), but the geometry of the initial collision is still ambiguous and several models have been proposed. Whereas some models favor an arc - continent collision (Huang et al., 2006; Suppe, 1984, 1988; Teng, 1990), others suggested a two stage collision of an exotic block with the Eurasian continental margin and a second collision of the Luzon volcanic arc with the passive margin (Lu and Hsü, 1992), or an arc - arc collision between the Luzon volcanic arc with a paleo - Ryukyu arc system extending to the west of Taiwan (Seno and Kawanishi, 2009; Sibuet and Hsu, 1997; Sibuet et al., 1995), or even that the collision may have happened synchronously along the entire margin length (Castelltort et al., 2011, Lee et al., 2015).

It is also known that the orogenic system in Taiwan reached an approximately constant mountain width of 90 km, which has been interpreted as being an expression of topographic steady state (Stolar et al., 2007; Suppe, 1981). Additional observations also show that the Western foredeep eventually reached a steady state size where accommodation space stayed constant despite the large sediment fluxes from Taiwan mountains (Covey, 1984; Covey, 1986). Therefore Covey (1986) suggested that sediment bypass out of the basin must have been an important factor that balanced accommodation space and sediment supply, maintaining the basin shallow marine, and preventing it from becoming overfilled or even fully terrestrial.

In this study, a 3D stratigraphic model is used to test different tectonic scenarios for the orogen evolution and how basin architecture corresponds. The model is calibrated with seismic lines from the Taiwan Strait. We show how different tectonic settings control the stratigraphic evolution of a foreland basin. This also involves a quantitative estimation of the sediment-volume budget for the basin and provides information on the importance of longitudinal sediment transport out of the basin.

2. General setting and background

2.1. Geology and Tectonics

The Taiwan mountains, rising almost 4 km above sea level, formed by the collision between the Philippine Sea plate and the Eurasian continent shelf. Arc volcanism associated with the subduction of the Philippine sea plate ceased between 6 Ma and 3 Ma, when the arc resisted subduction and collided with the Asian passive margin to form an initial accretionary wedge (Huang et al., 2006; Yang et al., 1995). The arc-continent collision is estimated to have initiated in late Pliocene (Nagel et al., 2012c). This is based on observing a continuous sandstone provenance shift and increasing illite crystallinity, interpreted to represent the progressive unroofing and recycling of the metamorphic orogenic belt (Dorsey et al., 1988; Nagel et al., 2012c). The oblique collision between the N-S trending Luzon volcanic arc and the NE-SW trending passive margin resulted in a southwest propagating collision (e.g., Nagel et al., 2012c; Simoes and Avouac, 2006; Suppe, 1981; Teng, 1990) and the modern collision point is presently located in the offshore SW Taiwan (Lin et al., 2008; Yu and Huang, 2009). The easternmost part of the South China Sea is currently being subducted below the Philippine sea plate along the Manila Trench whereas the Philippine sea plate itself is being subducted northwards below the Eurasian plate (Kao et al., 2000) together with high active seismicity and a convergence rate of 70 - 80 km/Ma between the Philippine sea plate and the Eurasian continent (Seno et al., 1993; Wu et al., 2009; Wu et al., 2007; Yu et al., 1997). The current plate convergence is mainly accommodated within the Longitudinal Valley Fault on the east coast and at the deformation front in the Western Foothills consistent with the main active faults (Yu et al., 1997).

The continental margin experienced extensive rifting and continental breakup phases due to the opening of the South China Sea in late Paleogene, which resulted in major subsidence and numerous sub-basins separated by topographic highs (Lee and Lawver, 1995; Lin et al., 2003). The orogen is divided into different tectonic units (Fig. 1) consisting of the accreted volcanic arc (e.g. Coastal Range) separated by the suture zone (e.g. Longitudinal Valley Fault), the main orogenic belt (e.g. Central Range), the deformed and uplifted foreland basin strata which constitutes a classical fold-and-thrust belt (e.g. Western Foothills), and the undeformed onshore (e.g. Coastal Plain) and offshore foreland basin sediments (Ho, 1988).

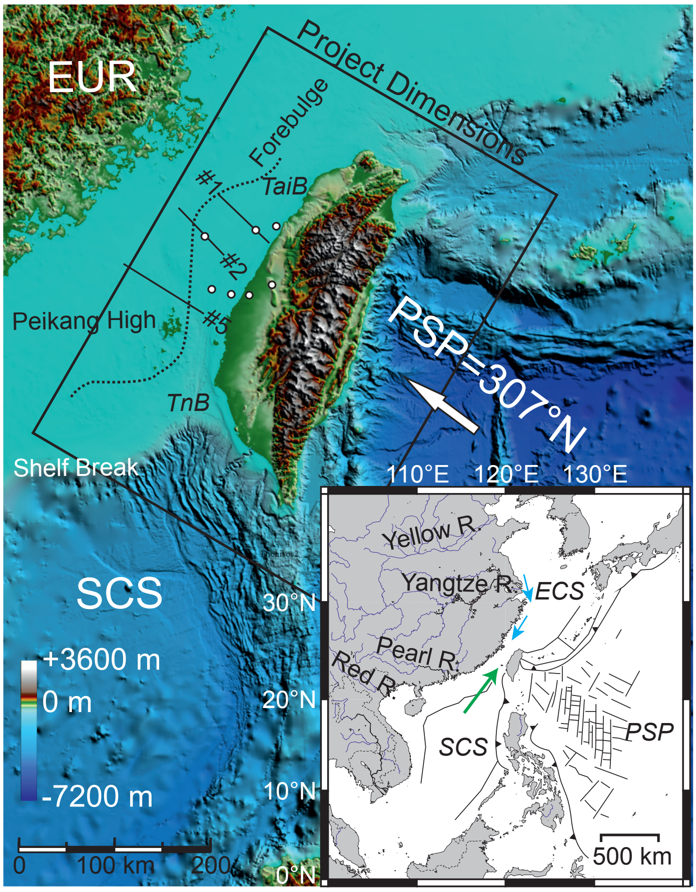

Figure 1: Bathymetric and plate tectonic framework of the studied area showing boundaries of stratigraphic model. Inset shows large-scale plate setting. Seismic lines tracks are from Yu and Chou (2001). Inset shows the four major Chinese river systems in the area of Taiwan. Large green arrow is South China Sea Current (northward), and small blue arrows are China Coastal Currents (southward). EUR: Eurasia. ECS: East China Sea. PSP: Philippines Sea Plate. SCS: South China Sea. TaiB: Taishi Basin. TnB: Tainan Basin.

As initially described by Covey (1984), the evolution of the syn-collisional facies is very similar from north to south, except for distinctive grain size contrasts (Chou, 1973). The coarse fraction was trapped in a shallow continental shelf basin (e.g. Taishi Basin), which was separated from the South by a topographic barrier (the "Peikang High", Meng, 1967). Most of the fine grain size was transported further southwards and became deposited in a deep marine basin to the South (e.g. Tainan Basin and/or South China Sea).

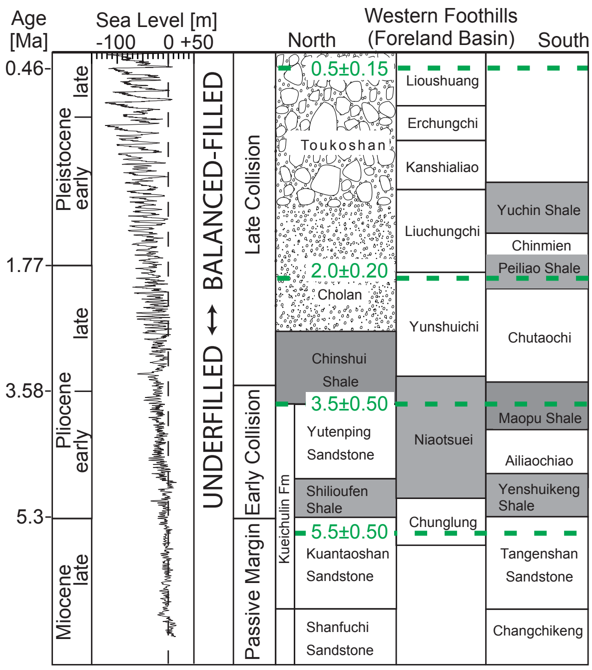

Figure 2: Schematic stratigraphy of the foreland basin of Taiwan from north to south. Note the change in lithology in Central and South-Taiwan due to the topographic barrier called “Peikang High”, which separated the northern Taishi basin and the southern Tainan basin. Thick Conglomeratic units up to 3'000 m are found in the Toukoshan fm., which is made of coarse alluvial and fluvial sediments. The sea level curve is adapted from Miller et al. (2011). The key dated horizons (Nannofossils, see synthesis in Nagel et al., 2013) are marked with thick dashed lines. Shales are in grey. Sandstones or sandstones and shales are in white. The Cholan formation is particularly coarse-grained, and the Toukoshan formation is conglomeratic.

The stratigraphic succession comprises a first retrogradational series consisting of shallow marine deltaic environments, which are often tidally influenced (Fig, 2, Kueichulin fm.). The transgression associated with the end of this formation marks the onset of orogenic loading of the shallow marine shelf environment. To the south, the formation passes progressively into deeper marine mud-dominated deposits (Fig. 2). The source of sediments during the deposition of the Kueichulin fm. is essentially the same as during the previous passive margin history of the basin, from the Eurasian continent to the southeast (Castelltort et al., 2011; Nagel et al., 2013; Shaw, 1996). It is followed by the Pliocene Chinshui Shale, a relatively deep marine mud-dominated formation, which constitutes the underfilled stage of the foreland basin (Covey, 1984). Reworked fossils, paleocurrent directions and facies analysis point to a main source from the east of the basin at this period, which is the growing orogenic wedge (Chang and Chi, 1983; Nagel et al., 2012a; Nagel et al., 2012c). The Cholan formation represents a large-scale progradational sequence of shallow marine wave- and tideinfluenced environments, which became progressively dominated by fluvial processes upsection. This is the main foreland basin stage driven by large sediment fluxes out of the Taiwan orogen and southward migration of facies belts. During the late Pleistocene, increased erosion lead to the deposition of large alluvial sediments which most likely are an ancient example of braided rivers draining the orogen today (Covey, 1984).

2.2. The foreland basin unconformity

The age of collision onset and its kinematics are still controversial. Most authors consider a collision age between 6.5 Ma and 3 Ma assuming a single arc-continent collision (e.g., Chang and Chi, 1983; Dorsey and Lundberg, 1988; Huang et al., 2006; Lin et al., 2003; Pelletier and Stephan, 1986; Suppe, 1981; Teng, 1990) or a two stage collision (e.g. arc-arc and arc-passive margin) between 12 Ma and 3Ma respectively (Lu and Hsü, 1992; Seno and Kawanishi, 2009; Sibuet and Hsu, 1997; Sibuet et al., 1995).

The flexural response due to the loading of the Eurasian shelf by the forming orogen and its sedimentary response has been studied in detail (Castelltort et al., 2011; Chen et al., 2001a; Chiang et al., 2004; Simoes and Avouac, 2006; Tensi et al., 2006). Tensi et al. (2006) suggested that the passive margin lithosphere already experienced flexure since 12.5 Ma and interpreted the observed flexure as not being related to the initial arc-continent collision, which is consistent with plate kinematic reconstructions (Hall, 1996; Nagel et al., 2012a; Sibuet and Hsu, 2004). The basal foreland unconformity is observed in the Northern basins (e.g. Taishi basin, Fig. 3) with an age estimated between 8.6 and 5.6 Ma (based on biostratigraphic data), consistent with a flexural migration of the load from east to west (Lin and Watts, 2002; Lin et al., 2003). This unconformity separates the passive margin sequence and the foreland basin sequence, which onlaps onto it. The depositional hiatus increases in duration from the current frontal thrust towards the forebulge in the middle of the Taiwan Strait (Lin et al., 2003; Yu and Chou, 2001).

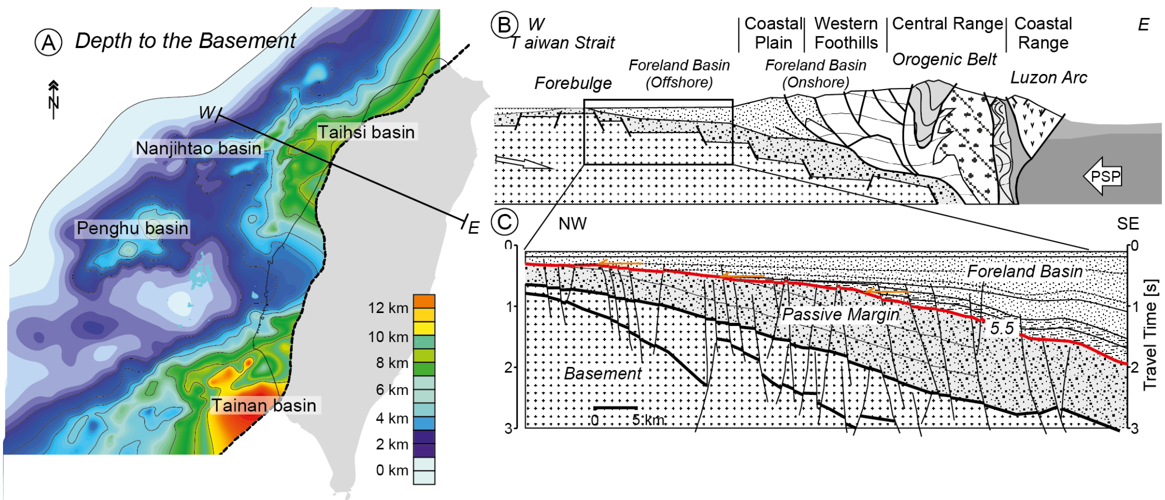

Figure 3: A) Map of depth to the Cenozoic basement In Taiwan's foreland, highlighting four individual basins separated by basement highs. The Nanjihtao basin is partially covered by a distal foreland basin sequence. The Taishi and Tainan basin are separated by the Peikang High. Figure modified from Lin and Watts (2003). B) Schematic cross section of the Taiwan orogen in the north, where the interpreted forebulge topography is most pronounced. C) Detailed seismic line drawing modified from Yu and Chou (2001). The foreland basin sequence onlaps onto passive margin deposits, with increasing depositional hiatus towards the forebulge, where late Pleistocene-Quaternary sediments directly overlie Miocene strata.

2.3. Modern sediment fluxes in Taiwan

The East Asian monsoonal climate was most probably established since 8.5 Ma (e.g. late Miocene) with an intensification observed since 5 to 3 Ma (Liu et al., 2003; Wan et al., 2006; Zheng et al., 2004). Today the island experiences 4 to 6 typhoons with maximum mean annual rainfall of 2000-3000 mm yr^{-1} (Kao and Milliman, 2008). The total annual amount of sediment delivered to the ocean by Taiwanese rivers has been estimated to be up to 500 Mtyr-1 with a strong asymmetry across the mountain range (Dadson et al., 2003; Liu et al., 2008). Estimates of erosion rates range between 2.2 to 8.3 mmyr^{-1} (Dadson et al., 2003; Fuller et al., 2003) and up to 30 mmyr^{-1} (Resentini et al., 2017) in agreement with quantitative estimates from thermochronometric constraints of 3 to 10 mmyr^{-1} (Lee et al., 2006; Willett et al., 2003).

Much of the suspended sediment is delivered at hyperpycnal concentrations into the Taiwan Strait (Dadson et al., 2005; Milliman and Kao, 2005; Milliman et al., 2007) where it is redistributed by seasonal and tidal currents (Jan et al., 2002). The northeast directed South China Sea current (Fig. 1), for example, transports warm tropical water into the Strait with a peak intensity during the summer month (June to August), in contrast to the southwest directed China Coastal current (Fig .1), which can deliver small portions of Yangtze-derived mud into the northern Taiwan Strait (Hu et al., 2010; Xu et al., 2009) during winter month (September to May). The morphology of the sea floor as well as the seasonal variation in temperature and salinity exert an important control on the distribution of grain size in the modern sediments and may help to understand the occurence of different facies in the ancient sedimentary record (Liao et al., 2008; Yang and Chun, 2001).

Marine observations indicate that fine mud particles are relatively quickly transported northward out of the Taiwan Strait (Horng and Huh, 2011; Horng et al., 2012; Huh et al., 2011; Liu et al., 2010). For example, marine investigations in the Choshui river delta made before and after a typhoon hit the island, showed that fine-size particles are redistributed and transported northward within a month (Milliman et al., 2007).

Average sedimentation rates vary greatly from 2 mm yr^{-1} in the Western foreland basin to 3-4 mmyr^{-1} in the Coastal Range (Chen et al., 2001a; Lin et al., 2003; Lundberg and Dorsey, 1990), with a rapid increase observed since the onset of deformation in the Western Foothills (Chang et al., 1983; Lock, 2007; Mouthereau and Lacombe, 2006; Mouthereau et al., 2001). These values are in accordance with erosion rates estimates of between 2 and 10 mm yr^{-1} from modern river sediment loads and interpretation of thermochronological data (Dadson et al., 2003; Fuller et al., 2003; Fuller et al., 2006; Liu et al., 2001; Liu et al., 2000; Siame et al., 2011; Simoes et al., 2007; Simoes and Avouac, 2006; Willett et al., 2003)(Table 1).

Author/Year	Uplift (U)/Exhumation Rate (E) [mm/a]	Erosion Rate/Incision Rate (I) [mm/a]
Peng et al. (1977)	U: 5±0.7 (<9 ka) Holocene coral reefs [1]	
Liu (1982)	U: 4.2-6.8 (3-0.5 Ma) 8.9±1.9 (<0.6 Ma)	
Jahn et al. (1986)	U: 3-4 (<3 Ma) Rb-Sr isotopes (TC)	
Lundberg and Dorsey (1990)	U: 5.9-7.5 *(<1.3-0.9 Ma)* CoR	6-7 (<1 Ma)
Wang and Burnett (1990)	U: 1.2-6.1 *10 ka* (Holocene) [2]	
Chen et al. (1991)	U: 5-14 *(<5000 a)* (CoR, uplifted corals)	
Liew et al. (1993)	U: 2.5-8 (Holocene) *elevated shoreline deposits (CoR)*	
Lo and Onstott (1995)	E: 1.7-1.6 *(K-Ar reset ages)*	
Liu (1995)	U: 36-42 *(<10a, GPS, CR)*	
Liu et al. (2000, 2001)		2.5-4.6 (<4 Ma) 2.3-6 (TC) ZFT/AFT
Hsieh and Knuepfer (2002)	U: <10 (Holocene river terrace)	I: <20 (river incision)
Dadson et al. (2003)	E: 3-6 (ECR) 1.5-2.5 (SW Taiwan)	5.2 (<30a) SSC [3] 6 (CR), up to 60 [4] I:1.5-9 (Holocene)
Fuller et al. (2003)		2.2-8.3 (8-27a, SSC)
Willett et al. (2003)		7-8 4-6 (AFT/ZFT)
Yamaguchi and Ota (2004)	U: 5-15 (<13 ka) Holocene marine terraces (CoR)	
Song et al. (2004)	U: 10.9, 5.4 Holocene marine terraces (CoR)	
Simoes and Avouac (2006); Simoes et al. (2007b,a) [5]		4.2 (BR)-6.3 (TC) 2-3 *(<1.5 Ma)*
Fuller et al. (2006)	E: 3-5 *(acceleration since 2-1 Ma)*	2.3-3.3 (AFT/ZFT [6]) max. 6 - 8
Lee et al. (2006)	U: <1 *(6-1 Ma)* 4-10 *(<1 Ma)*	
Siame et al. (2011)		2 ±1 *(<100 ka)*, 5-7 (<50 a) [7]
Kuo-En (2011)	U: 0.2-18.5 *(2000-2008, GPS)* (BR)	
Siame et al. (2012)		I: 0.8±0.1 - 10.1±1.3 (<300 ka) (Choshui river terraces)

[1]Hengchun Peninsula, Tainan area, Coastal Range
[2]Hengchun Peninsula, Coastal Range, Lanyu and Lutao
[3]SSC=calculation based on modern suspended sediment concentrations
[4]active thrust faults, Western Foothills, Southwest Taiwan
[5]thermokinematic modeling
[5]BR=Backbone Range, TC=Tananao complex, CR=Central Range, ECR=Eastern Central Range, CoR=Coastal Range
[6]AFT=Apatite fission track, ZFT=Zircon fission track
[7]Be10, Lanyang catchment

Table 1: Summary of observed and predicted kinematic data for the Taiwan orogen.

3. Data sets and methods

3.1. 3D stratigraphic model "Dionisos"

To evaluate the complex relationships between the stratigraphic record, tectonics (e.g. subsidence, with respect to the initial collisional geometry) and climate (e.g. erosion rates), the stratigraphic model Dionisos was used (Granjeon, 1997). Dionisos is a process-based modelling tool using a diffusion and advection law that links sediment flux to local slope (e.g. potential available energy to move sediment) and water flow (e.g. transport efficiency of the lithologies defined) by a diffusion coefficient. Erosion and sedimentation at each point of the basin are defined by combining the transport equation and the law of mass conservation:

$$Q_{sed} = -K \cdot Q_{water} \cdot \overrightarrow{grad}\, h$$

The second basic assumption of the model is the law of mass conservation

$$\frac{\partial h}{\partial t} = -div\, Q_{sed}$$

where:

Q_{sed} = sediment transport [m^2/s]

Q_{water} = relative water flow [-]

K = diffusion coefficient [m^2/yr]

h = ground elevation [m]

$\delta h/\delta x$ = elevation gradient (i.e., slope)

Boundary supplies (i.e. sediment volume and sand, mud fraction), water discharge of rivers at source locations and rainfall are defined for each sedimentary sequence. It is important to note that all the water introduced by the rivers and rainfall is conserved and flows towards the lowest part of the basin (Granjeon and Joseph, 1999). The potential sediment availability is simulated by a maximum erosion rate, which depends on climate (e.g. rainfall), subsidence rate and uplift rate (e.g. topographic elevation).

The study area was set as a 500 km x 320 km rectangle in the Taiwan Strait where abundant data is available (Fig. 1). It is confined to the flexural forebulge in the West and the Coastal Range in the East, and includes the Taishi basin in the North and the Tainan Basin in the South (Fig. 1). The main input data required by Dionisos consist of tectonic subsidence for different time intervals, sediment supply and eustatic sea level fluctuations (e.g. climatic influence), compaction, flexure and sediment transport parameters. Sediment influx corresponds either to a predefined boundary condition into or out of the study area or to basement erosion.

The input data was acquired from published boreholes and seismic lines offshore in the Taiwan Strait and onshore (Lin and Watts, 2002; Lin et al., 2003; Yu and Chou, 2001), together with constructed depth maps (Fig. 4) for five key stratigraphic horizons defined in an earlier study (Nagel et al., 2012c) provide a solid first approximation database.

3.2. Foreland basin subsidence

The most important basin-scale controls on accommodation include flexural tectonics related to tectonic loads and sea level changes. The West Taiwan basin formed by flexural bending of the Asian passive margin in front of the westward migrating thrust loads of the growing accretionary wedge (Lin et al., 2003). In order to better constrain the subsidence of the sedimentary basin, backstripping techniques (using Airy isostasy) were applied to 28 boreholes and 9 stratigraphic sections (Nagel et al., 2012c; Watts and Ryan, 1976).

Backstripping is used to stepwise decompact and unload a borehole or stratigraphic section from the influence of water and sediments and, therefore, to isolate the contribution of the tectonic forces responsible for subsidence. The tectonically driven subsidence at any location in the basin is given in Allen and Allen (2009):

$$TS = Y \cdot \left(\frac{\rho m - \rho s}{\rho m - \rho w}\right) - \Delta sl \cdot \left(\frac{\rho w}{\rho m - \rho w}\right) + (W_d - \Delta sl)$$

where:

W_d = the average water depth at which the sedimentary units were deposited
Y = decompacted sediment thickness
$\rho m/\rho w/\rho s$ = densities of the mantle, the water and mean sediment density
Δsl = the difference in sea-level height h between the present and the time at which the sediments were deposited:

$$\Delta sl = \left(\frac{\rho m - \rho w}{\rho m}\right) \cdot (h2 - h1)$$

The water depth at the time of deposition for the backstripped strata was estimated by applying the depositional model constructed by Nagel et al.(2012c). Note that since the sediments in the western foreland basin were deposited on a shallow marine continental shelf, the influence of the water column (10s of metres) on the backstripped strata is small relative to the considered thicknesses (100s of metres). Sediment was assumed to be composed of two main grain size classes, sand and mud, which correspond to the modern siliciclastic river supply and is consistent with detailed lithologic analysis (Huh et al., 2011; Nagel et al., 2012c). When the basin gets progressively filled with sediments, mechanical compaction introduces loss of water during sediment burial and affects the depth-porosity curves for different lithologies. The trend between porosity and depth is usually approximated by:

$$\emptyset = \emptyset_0 \cdot e^{-cy}$$

This produces an asymptotically low porosity with increasing depth, where $\emptyset_0$ describes the surface porosity and c the coefficient of compaction (Table 2). The flexure of the basement was computed with an elastic thickness of 15 km, a Young's modulus of 100 GPa and a Poisson's ratio of 0.25. These values are in agreement with recently published values for the Taiwan foreland basin (Lin and Watts, 2002).

Lithology	Surface porosity $[\phi_0]$	Compaction coefficient $[km^{-1}]$	Density $[kg/m^{-3}]$
Shales	0.63	0.51	2720
Sandstones	0.49	0.27	2650
Mudstones	0.56	0.39	2680
Water			1030
Mantle			3330

Table 2: Values used in the backstripping for the different lithologies observed in the Western Taiwan foreland basin and compaction coefficient (after (Allen and Allen, 2009; Lin et al., 2003; Tensi et al., 2006).

The backstripping results provide a detailed record of the Asian passive margin subsidence and uplift history at five key biostratigraphic horizons (Fig. 4). Lin and Watts (2003) showed that the subsidence history of the Asian passive margin is strongly influenced by its syn- and post-rift history due to the extension in the South China Sea (e.g. post-breakup extension from 30 to 21 Ma, thermal subsidence from 21 to 12.5 Ma and a second post-breakup extension from 12.5 to 6.5 Ma). The increased subsidence since the early Pliocene is ascribed to the growth of the Taiwan orogen as it propagates westward, introducing deformation and increasing sedimentation rates in the basin (Chang and Chi, 1983; Mouthereau et al., 2001). In addition, Tensi et al. (2006) demonstrated that the load associated with the initial foreland basin has migrated rapidly westward 1 Ma ago and was stabilized at the same time as the basin was buried under large quantities of sediments (e.g. alluvial and fluvial fans of the Toukoshan fm.).

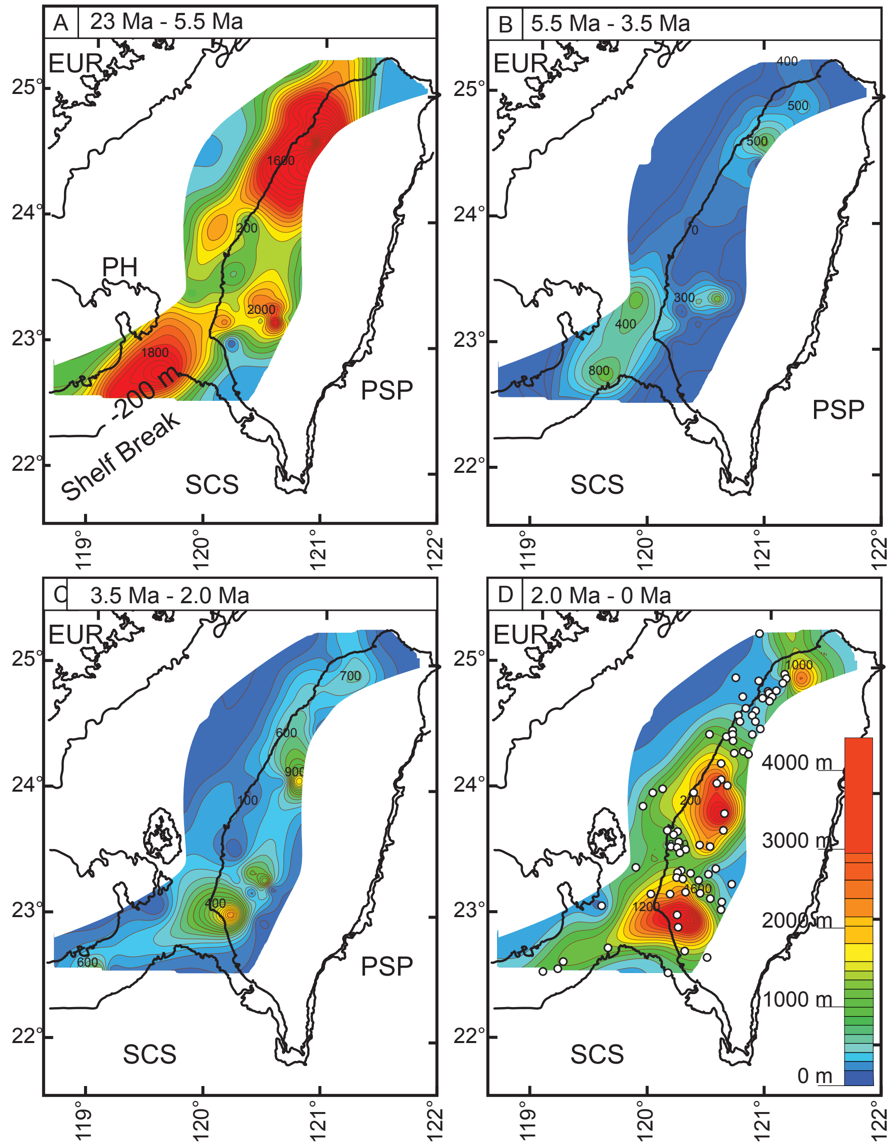

Figure 4: Maps of decompacted sediment thickness in between the five key biostratigraphic (nannofossils) horizons of Nagel et al (2013).

The reconstructed subsidence pattern is consistent with sediment isopach maps shown in Figure 4. Lin and Watts (2002) showed that the topography is insufficiently high to produce the observed subsidence pattern in an isostatic flexural model driven by surface loads. Following Simpson (2014), it can be proposed that this observation is an illustration of a possible decoupling between subsidence and surface loads, especially prominent in deeply eroded mountain ranges. In his model, Simpson explains that what may have been previously attributed to “buried loads” (as in Taiwan, e.g. Lin and Watts, 2002) could be related to the accumulation though time of vertical deformation due to repeated large seismic events and the dragging of the foreland margin by reverse slip on the main orogenic front.

3.3. Sediment fluxes and basin boundaries

The volume of sediment deposited in the basin was calculated from data of published boreholes drilled by the CPC (Chinese Petroleum Corporation) in the western foreland basin (Fig. 4) (e.g., Lin et al., 2003; Shaw, 1996). For each sequence the sediment thickness is extrapolated between the present day forebulge and the Western Foothills by a triangulation algorithm to obtain four maps between the early Miocene and late Pleistocene. The chronostratigraphy of the Western Foothills has been extensively studied with Neogene calcareous nannofossils (Chang and Chi, 1983; Chou, 1973; Huang, 1977; Huang and Huang, 1984) and builds the basis for the five key biostratigraphic horizons that are best documented (Nagel et al., 2013). Two main depositional basins are recognized throughout the Neogene: the Tainan basin in the South and the Taishi basin in the North (Fig. 3). During most of the Miocene, sediment accumulation rates were low, but started to increase in late Miocene to early Pliocene (Chang and Chi, 1983). The early Pliocene shallow-water Kueichulin formation is generally associated with the latest pre-orogenic deposition on the passive margin and the relative sea level rise recorded in the transition from Kueichulin to Chinshui formations marks the onset of load induced subsidence by the growing accretionary wedge (Fig. 2). During the late Pliocene, the mud-dominated Chinshui Shale was deposited into an underfilled, but relatively shallow marine foreland basin (Covey, 1984), which was then filled by the nearby orogenic wedge with fluvial and alluvial sediments during Toukoshan fm. It is important to note that in the modern Taiwan Strait, mud-sized grains are quickly transported northwards out of the Taiwan Strait (Milliman et al., 2007, Liu et al., 2008, Kao et al., 2008a, Huh et al., 2011) and that the sediments eroded from the orogen possibly contained a larger amount of mud than currently found in deposits, and that has been fractionated away by marine processes.

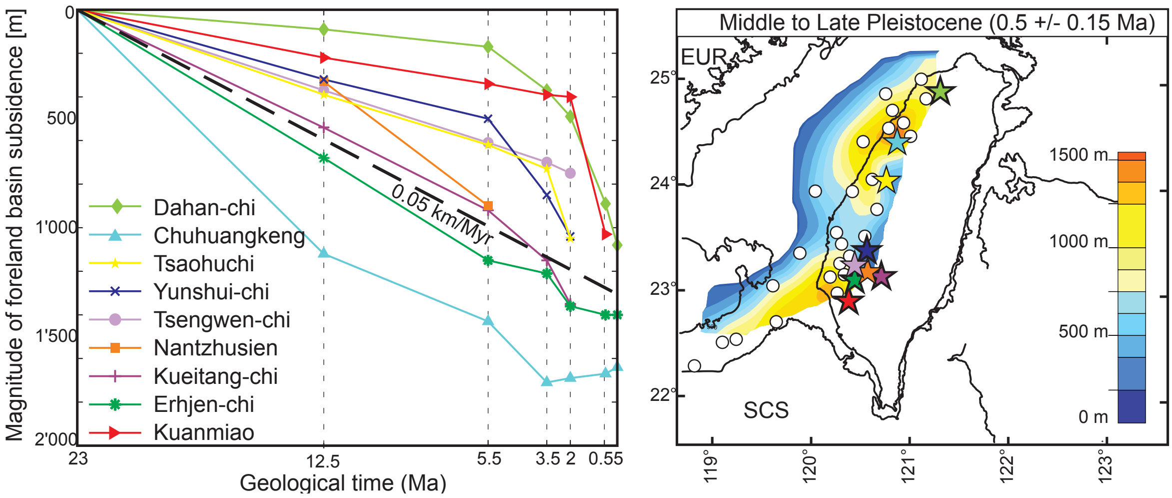

Figure 5: A) Tectonic subsidence histories from stratigraphic sections in the fold-and-thrust belt of Taiwan (marked with a star, ordered from north to south). Dahan-chi (Pan, 2011), Chuhuangkeng (Huang, 1976), Yunshuichi/Tsaohuchi (Yeh and Chang, 1991; Yeh and Yang, 1994), Tsengwen-chi (Chen et al., 2001a), Nantzhusien (Ting et al., 1991; Yu et al., 2008), Kueitangchi (Huang, 1977), Erhjen-chi (Horng and Shea, 1994), Kuanmiao (Chiu, 1975). B) Tectonic subsidence map for the late Pleistocene (NN19/20) with 28 boreholes and 9 stratigraphic sections. Stars indicate stratigraphic sections with color coding corresponding with left panel in which sections legend is ordered from North (top) to South (bottom).

The sediment volume accumulated within each time sequence is shown in Table 3. A total of 82-125'000 km^3 of sediment accumulated since 5.5 Ma in the foreland basin of Taiwan. If we assume that the collision started between 5.5 Ma and 3.5 Ma, and that before 5.5 Ma the sediment thickness corresponded only to the influx of material from Asia mainland, we interpret the increasing sediment influx from the Taiwan orogenic wedge to be in the range of 6'500 to 28'000 km^3/Ma (Table 3). This sediment flux estimate is probably overestimated since the basin also must have received material from its western border, i.e. Asia mainland. However, this contribution was swamped by the dramatic increase in sedimentation rates that accompanied Taiwan orogeny (Chang and Chi, 1983). In addition, some of the sediment transported into the foredeep consists of recycled foreland basin deposits. Therefore the calculated sedimentation rates over the area of the modern foreland basin are lower when compared with sedimentation rates from the Western Foothills, especially during the last phase of orogenesis from 2 Ma to 0 Ma (Chang and Chi, 1983).

Age [Ma]	Sediment Volume [km^3]	Sediment Flux [km^3/Ma]	Sediment Flux [km^3/Ma] From Asia	Sediment Flux [km^3/Ma] From Taiwan
0.0 - 2.0	47'214 - 64'884	23'607 - 32'442	¨	17'864 - 28'203
2.0 - 3.5	20'443 - 35'980	13'628 - 23'987	¨	7'885 - 19'748
Total	*67'657 - 100'864*			
3.5 - 5.5	15'917 - 26'502	7'959 - 13'251	¨	6'594 - 9'012
5.5 - 23.5	76'298 - 103'383	4'239 - 5'743	4'239 - 5'743	0
Total	*159'872 - 230'749*			

Sediment Volumes SE Asia	[km^3]	2 - 0 Ma	2 - 5 Ma	5 - 11 Ma	11 - 17 Ma	17 - 24 Ma	24 - 5 Ma	[km^3/Ma]
Pearl River & S. Taiwan	Total	81000	93000	240000	85000	130000	455000	23947
	uncertainity	*40000*	*38000*	*100000*	*31000*	*49000*		
	min	61000	74000	190000	69500	105500	365000	19211
	max	101000	112000	290000	100500	154500	545000	28684
E. China Sea & N. Taiwan	Total	190000	200000	45000	39000	88000	172000	9053
	uncertainity	*81000*	*75000*	*16000*	*12000*	*31000*		
	min	149500	162500	37000	33000	72500	142500	7500
	max	230500	237500	53000	45000	103500	201500	10605
Okinawa Trough	Total	130000	99000	16000	5200	4900	26100	1374
	uncertainity	*64000*	*40000*	*6000*	*2000*	*2100*		
	min	98000	79000	13000	4200	3850	21050	1108
	max	162000	119000	19000	6200	5950	31150	1639
Total	Total	271000	293000	285000	124000	218000	653100	34374
	Min	210500	236500	227000	102500	178000	528550	27818
	Max	331500	349500	343000	145500	258000	777650	40929

Table 3: Sediment volume accumulated during the Neogene on the Asian passive margin calculated for the area between the modern forebulge and the Western Foothills (ca. 35'000 km^2). The age sub-division corresponds to biostratigraphic key horizons from nannofossil zonation (Nagel et al., 2013). The first two digits are considered significant.

Table 4: Sediment volume accumulated in the Cenozoic sedimentary basins of Southeast Asia (modified from, Métivier et al., 1999).

Sediment fluxes from the growing Taiwan orogen are delineated by comparing the amount of sediment that has been preserved in the Taiwan foreland basin with the amount of sediment accumulated in the Cenozoic sedimentary basins north and south of Taiwan (Table 4). The obtained boundary supply fluxes are minimum, because some unknown amount might have bypassed or not even reached the Taiwan Strait.

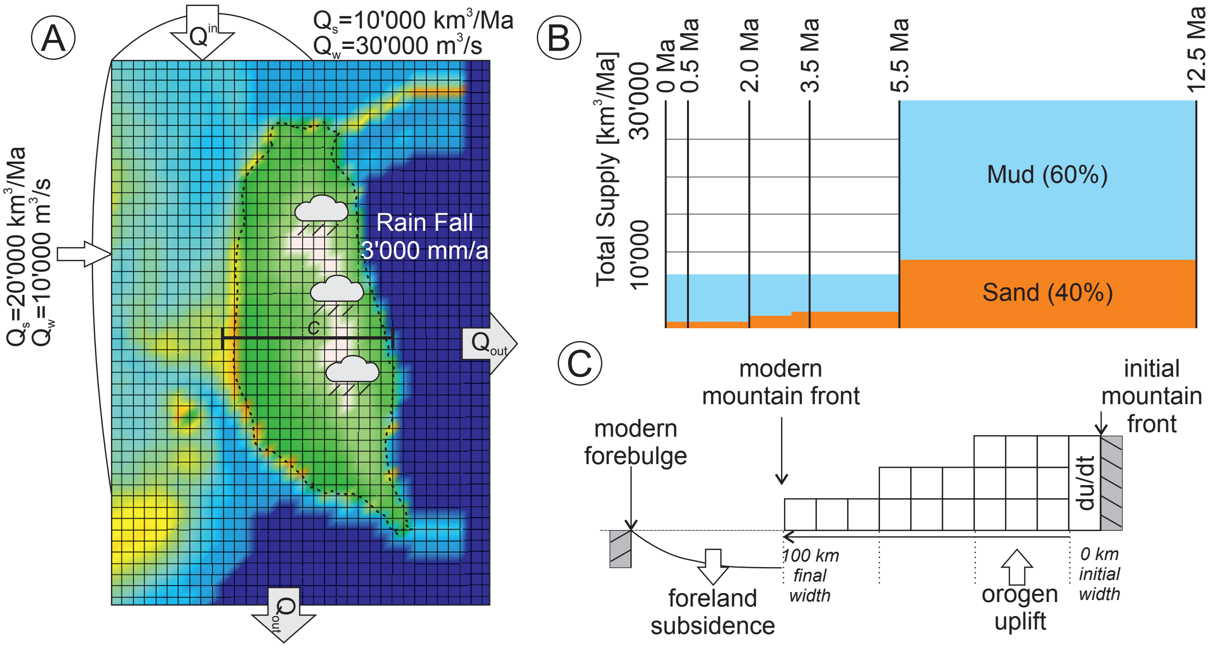

Figure 6: Initial sediment supply history applied in the model. A) The two source areas correspond to Asia mainland (West) and the East China Sea (North). Rainfall follows the modern mean annual rainfall rate. The fluvial water discharge was estimated by modern river discharges in Southeast Asia (Table 5). B) The initial total sediment supply from the boundaries through time is based on the sediment preserved in the foreland basin and the sediment accumulation rates in Southeast Asia (Métivier et al., 1999). The grain size distribution follows modern suspended sediment concentrations and sea surface measurements in the Taiwan Strait (Huh et al., 2011; Kao and Milliman, 2008; Xu et al., 2009), as well as observations in the ancient sedimentary record (Nagel et al., 2013). See text for discussion. C) Sketch of orogen growth as implemented in Dionisos: the area of uplift rate is progressively enlarged to simulate mountain range widening, from 0 at the onset to 100 km width at the end.

Finally, the initial sediment supply history applied in the model is shown in Figure 6. Two source areas are defined along the western and northern model area (Fig. 1). It is important to note that these sources refer to general provenances located along the model boundaries and are not meant to represent individual rivers. Today, only the smaller tributaries of the Minjiang and Jiulong rivers drain directly into the Taiwan Strait, collectively discharging only 1/10 of the Taiwanese rivers (Table 5). The water discharge per source area was assumed to be similar to the modern water discharge of rivers in Southeast Asia (Table 5).

	catchment area [km^2]	mean channel gradient [m/km]	average water discharge [m^3/s]	sediment load [km^3/a]	shelf gradient [m/km]
WORLD*					
Mississippi	2900000	0.5	17704	0.1481	8.0
Amazon	5700000	0.8	150000	0.4444	0.7
Ebro	85000	1.3	500	0.0074	3
Nile	4000000	1.6	2700	0.0889	4
Bengal	1750000	1.7	29700	0.3630	1.1
Indus	1400000	2.5	2644	0.1667	1.5
Sepik River	77700	0.05	3700	0.0288	6.50
Fly River	76000	-	6000	0.0315	0.76
Waiapu River	1734	-	1346	0.0181	5.00
SOUTHEAST ASIA**					
Yangtze	1940000	0.04	28507	0.1778	0.17
Red	124400	2	1200	0.0370	0.39
Pearl	440000	-	10654	0.0167	0.52
Yellow	750000	8.2	1160	0.4000	0.15
Minjiang	61000		6000	0.0028	
Jiulong	14700		1000	0.0009	
TAIWAN***					
Touchien	566	0.35	25	0.0078	2.3
Houlung	536	0.44	22	0.0181	1.8
Taan	758	0.34	36	0.0007	1.8
Tachia	1235	0.26	78	0.0011	1.4
Wu	2026	0.22	116	0.0019	0.8
Choshui	3155	0.18	120	0.0004	1.4
Peikang	645	0.06	27	0.0037	4.1
Pachang	475	0.24	23	0.0222	4.3
Tsengwen	1177	0.18	34	0.0007	4.3
Erjen	350	0.07	19	0.0007	5.0
Kaoping	3256	0.23	248	0.0059	5.0
Peinan	1603	0.44	97	0.0078	30.0
Hsiukuluan	1790	0.29	125	0.0130	50.0
Hualien	1507	0.4	100	0.0333	35.0
Hoping	556	0.37	37	0.0081	80.0
Lanyang	979	0.48	75	0.0111	13.0

*from Somme et al. 2009 and references therein, **Chen et al. 2001, Wang et al. 2005, Wang et al. 2007, Wang et al. 2008, Liu et al. 2007,Zhang et al. 2008, Olariu et al. 2009, Milliman and Syvitski 1992, Yu et al. 2006, Kineke et al. 2000, Huh et al. 2011, Kuehl et al. 2004, Kuehl et al. 2000, Yu et al. 1991, Wolanski et al. 1995, Kniskern et al. 2010, Hicks et al. 2000, ***Liu et al. 2008, Kao and Milliman 2008, Yu and Chiang 1997

Table 5: Main parameters of rivers from Taiwan, Southeast Asia and larger rivers worldwide for comparison. Data World from Sømme et al,(2009) and references therein. Data for Southeast Asia from Chen et al (2001b), Hicks et al (2000), Huh et al (2011), Kineke et al (2000), Kniskern et al (2010), Kuehl et al (2004), Liu et al (2007), Milliman and Syvitski (1992), Olariu and Steel (2009), Wang et al (2007), Wang et al (2005), Wolanski et al (1995), Yu et al (1991), Yu and Huang (2006) and Zhang et al (2008), Data for Taiwan from Kao et al (2008), Kao and Milliman (2008), Liu et al (2008) and Yu et al (1997).

Since sediment transport in Dionisos is modeled by diffusion, a short review of published values for the diffusivity coefficient K in different depositional environments is provided here for comparison (Table 6). Although these cited modeling studies did not necessarily use diffusion in exactly the same manner(for instance depending on whether water discharge is taken into consideration or not), an average value for each depositional environment was used based on the values compiled in Table 6.

	Continental	Marine
Csato et al. (2007)	2000-4000	0.4-10
Clark et al. (2009)	1000-2000	0.01-1
Burgess et al. (2006)	125-500	2.5-10.0
Schlager and Adams (2001)	100	0.15
Flemings and Jordan (1989)	1-25.0	0.1-5
Jordan and Flemings (1991)	4-200	0.1-1
Sinclair et al. (1991)		0.5
Kaufman et al. (1991)		10.0-75
Rivenaes (1992)		0.05-10.0
Paola et al. (1992)	10.0-70.0	
Kenyon and Turcotte (1985)	24.0-560	
Naden et al. (1999)	191.0-3557	
Marr et al. (2000)	10-100	
Begin (1988)	4300	
Humphrey and Heller (1995)	0.25	

Table 6 Ranges of diffusion coefficients used in modeling studies for individual depositional environments. The values were converted to [km^2/ka].

Figure 7 shows the sensitivity of different parameters (water discharge, sediment thickness, sediment volume, sedimentation rate) for 7 model runs with increasing sediment transport efficiencies (between $0.1xK_{initial}$ and $1000xK_{initial}$). All the models were run with the standard model set-up described above. The parameters were measured at three different points within the basin at seismic line L1-1, L1-2, L5-1 (see Fig. 1) as well as the average value from 3.5 Ma to 0 Ma (marked with asterisk).

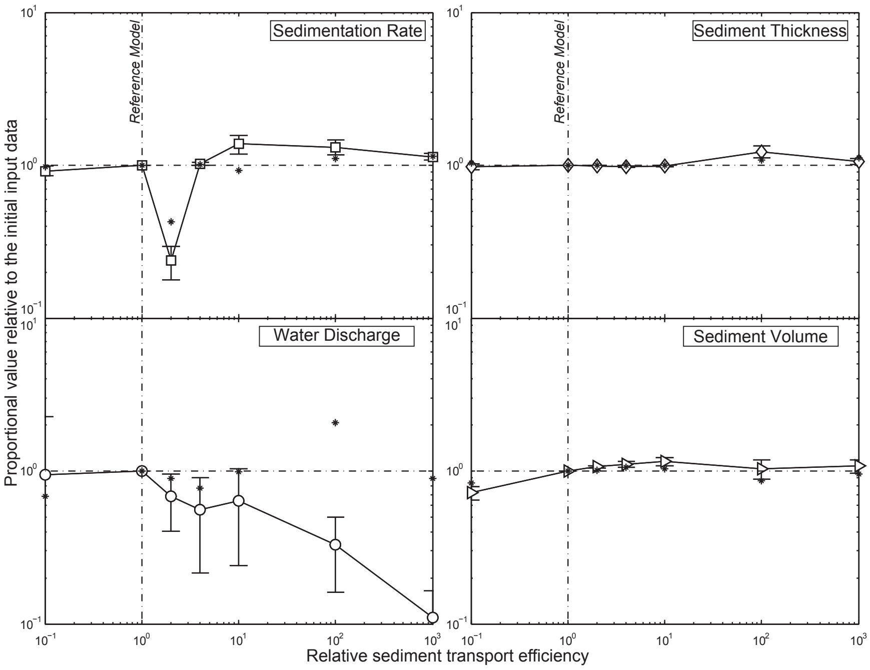

Figure 7: Sensitivity test for different diffusion coefficients. Proportional values of the four measured parameters in relation to the initial model input data within the foreland basin, plotted against the relative efficiency of sediment transport. The results are from seven separate model runs with the same initial reference model parameters, but different diffusion coefficients.

The model starts at 12.5 Ma, which corresponds to the NN5-6 nannofossil boundary. This key biostratigraphic horizons have already been used in an earlier study to reconstruct the paleogeography during the arc-continent collision (Nagel et al., 2013). The study shows that the sedimentation in the foreland basin during the Miocene to Pleistocene took place in a mixed storm- and tide-dominated shallow marine depositional environment. The paleobathymetry did not change significantly from 12.5 Ma to 3.5 Ma (Fig. 8), when the basin started to subside due to the approaching orogenic wedge in the east and the mud-dominated Chinshui Shale was deposited (Fig. 2). It is important to note that progradation and shallowing-upward cycles associated with the approaching orogenic wedge took place earlier in the northern parts of the basin and progressed southward as the basin was filling up (Nagel et al., 2012c).

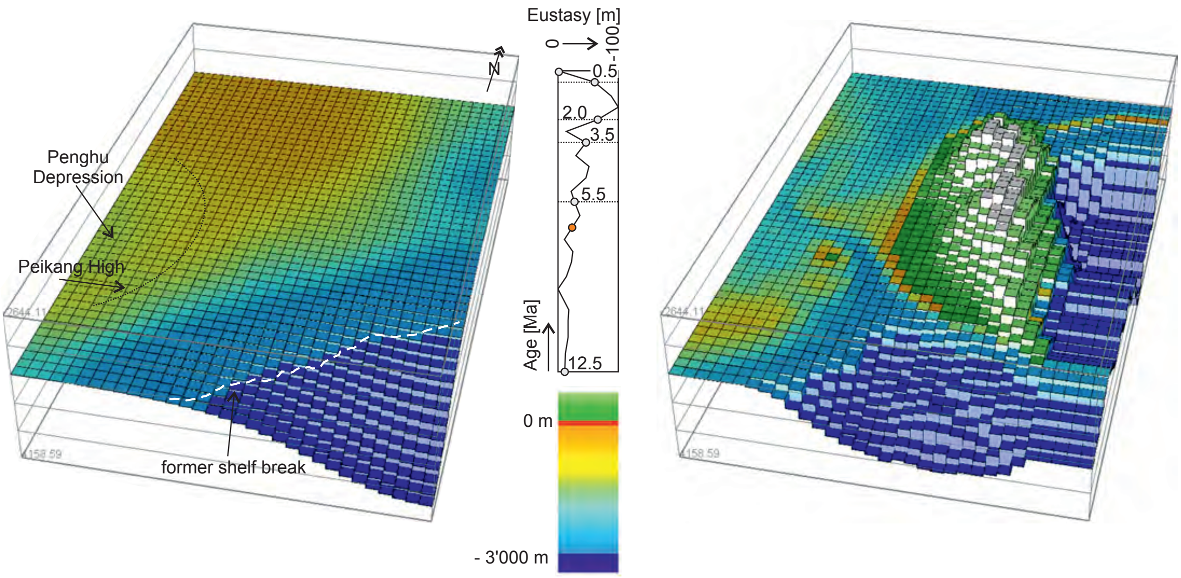

Figure 8: Initial bathymetry map at 12.5 Ma (left), and modern bathymetry (right). The 12.5 Ma map was constructed based on detailed facies analysis in the Western Foothills (Nagel et al., 2013). Sea level variation after Miller et al. (2011) for indication.

3.4. Experimental setup

To explore the orogen growth history and basin architecture, three different absolute tectonic scenarios were tested (Fig. 9). Each model considers the same initial boundary supply data (Fig. 6). In these experiments, orogenic uplift begins at ~4 Ma, which is in agreement with recent provenance studies (Nagel et al., 2013). The first orogenic growth model (Figure 9A) considers southward propagation of the orogen at a rate of 90 km/Ma until the present day length of 360 km is reached, and assumes a fixed steady state width of 90 km (Suppe, 1981). Using the time-space principle initially constructed by Suppe (1981), steady state size was reached after *ca* 1.3 Ma following the onset of orogeny. In a second model (Figure 9B) it is assumed that the orogen collided with a large promontory simultaneously along the length of the modern orogen, with no (or just minor) southward propagation. This scenario is based on sedimentological studies and paleogeographic reconstruction of Castelltort et al (2011) and tectonic-thermochronometric data of Lee et al (2015). The third model intermediate between both previous ones, (Figure 9C) considers a linear growth in length of the orogen with time, along with lateral displacement of the orogen area as it overthrusts the Eurasian margin. In all three models, a continuous and constant uplift rate of 5 km/Ma was assumed. This rate covers the range of uplift rates published in Taiwan (Table 1).

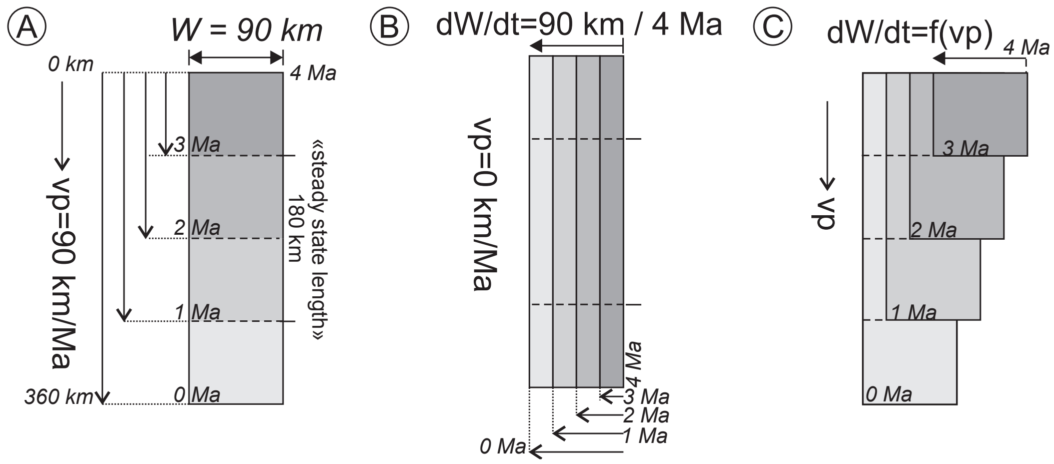

Figure 9: Three orogen growth models tested in this study. A) Pure lengthening: southward propagation (90 km/Ma) of a steady state orogen with a fixed width of 90 km. B) Pure widening: lateral propagation, with a fixed length of 360 km. C) Lengthening and overthrusting: southwestward propagation of a steady state orogen.

4. Simulation results and discussion

4.1. Foreland basin geometry

THE standard model setup needed to test the different tectonic scenarios (Fig. 9) is achieved by inposing subsidence maps for the different time intervals considered. The subsidence maps were constructed by first synthesizing the facies observed in the field into paleogeographic maps. These maps then provide an estimation of paleobathymetries for the entire basin (Nagel et al., 2013). Finally, the paleobathymetric estimates along with ages and sediment thicknesses are used to backstrip vertical sections in the basin, producing the subsidence maps (Fig. 10). An initial test of this approach is to try to recover the first order geometry observed on seismic lines in the Taiwan Strait (Fig. 3). A key horizon to compare is the transition from passive margin sedimentation to foreland basin sedimentation with its so-called "flexural forebulge unconformity". As shown in Figure 11, the imposed timing and subsidence results in sediment fluxes that correlate well with the observed geometry.

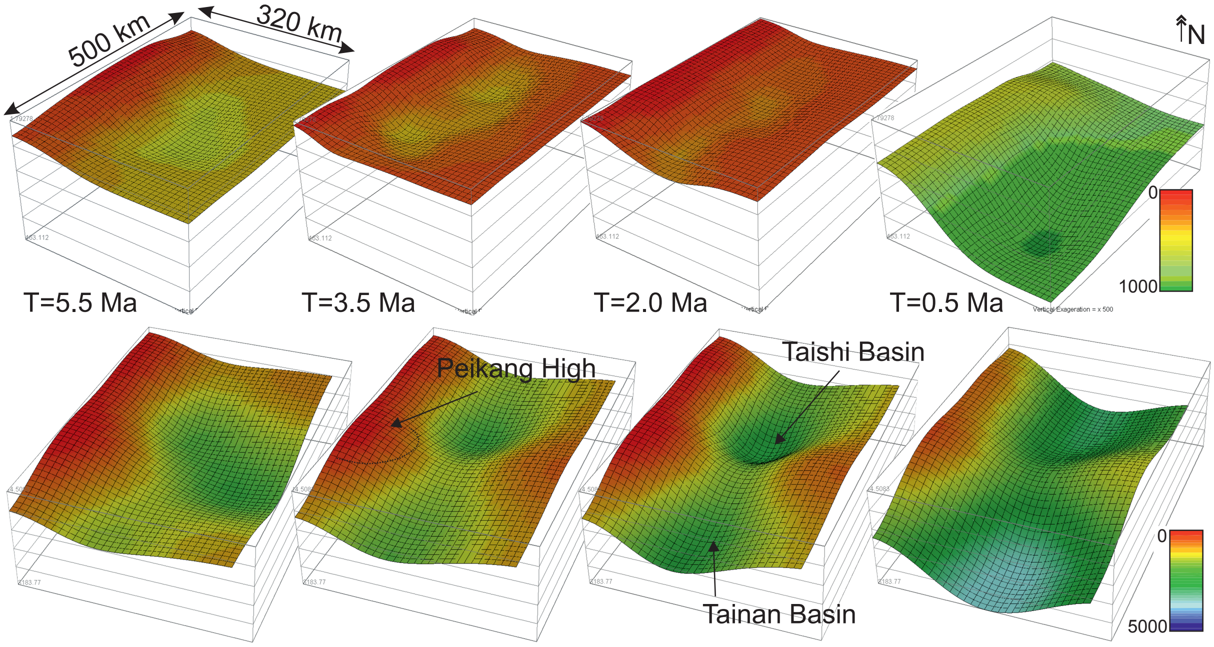

Figure 10: Smoothed subsidence maps for each simulated interval with tectonic subsidence used in the program (upper row) and total subsidence (including isostatic effects of sediment and water loading, lower row). The sediment depocenters in the north (e.g. Taishi basin) and the south (e.g. Tainan basin) are separated by the Peikang High.

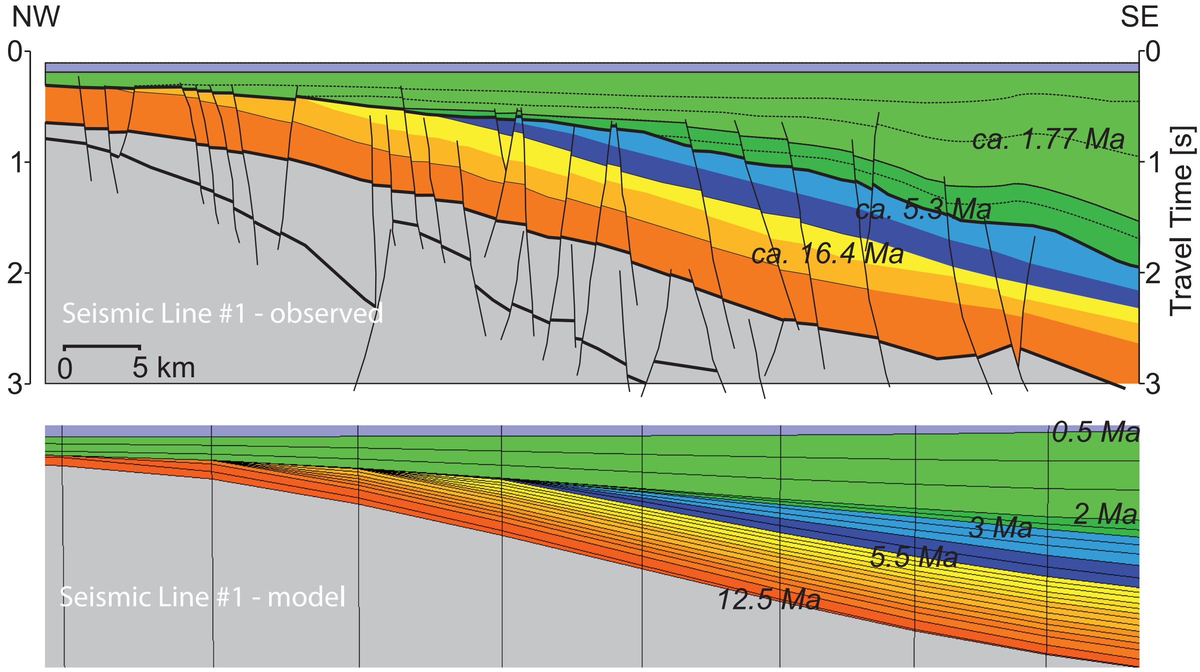

Figure 11: The input subsidence forces a dramatic change of sedimentation pattern at the transition from passive margin sedimentation to foreland sedimentation. This mimics the "flexural forebulge unconformity" documented by Yu and Chou (2001). This unconformity represents the boundary between the pre-collisional Nanchuang Fm. and the syn-collisional Kueichulin Fm. and was estimated approximately at 6.5 Ma (Lin et al., 2003).

4.1. Mass flux calculations

Theoretical models of mountain building propose that an orogen can reach a topographic steady state when the rates of rock uplift and erosion are balanced (Willett and Brandon, 2002). These models predict that, once steady state is reached, the sediment influx into the basin exceeds the available accommodation space, since no additional tectonic load is acting on the subsidence anymore, and therefore the basin becomes overfilled with time (Covey, 1986; Naylor and Sinclair, 2008). Despite observations suggesting that Taiwan has been in steady state since the Late Pliocene (Suppe, 1981, 1984), or even increased in exhumation rate in the Pleistocene (Hsu et al., 2016), the Western Foreland basin is still not overfilled. This can be explained either by a large original accommodation space or a continuous removal of sediment from the basin preventing it to fill-up.

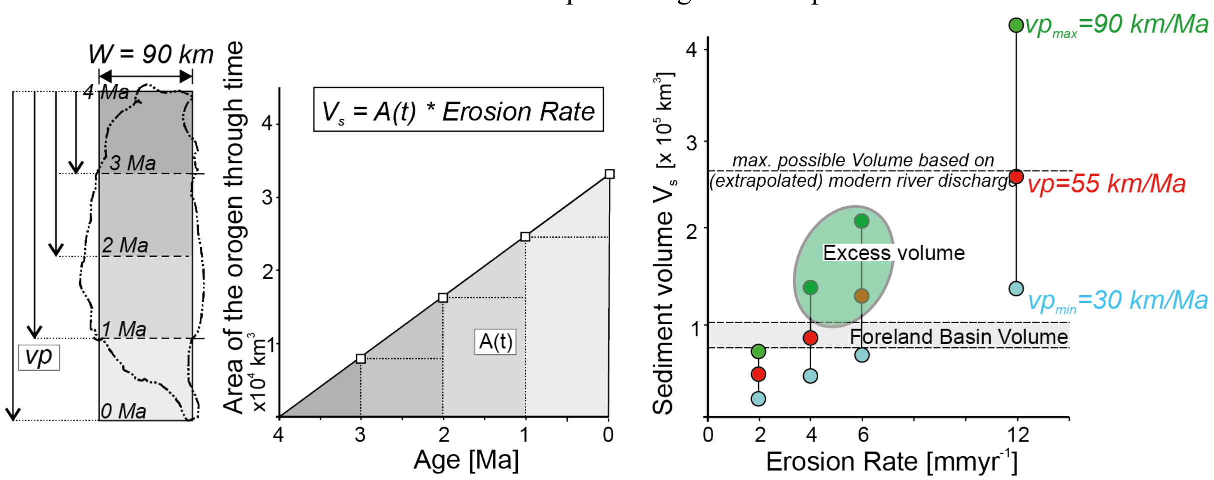

Figure 12: A) Orogen growth model with a steady state orogen width of 90 km and a southward propagation rate of Vp=30 to 90 km/Ma. B) The theoretical volume eroded from the mountain was calculated by integrating the orogen area through time multiplied by the erosion rate. C) The modern river discharge was extrapolated over 3.5 Ma and taken as an upper limit for the maximum possible sediment influx into the basin system (Table 5). The theoretical sediment volume eroded from the mountains was corrected for the fluvial discharge flowing to the east, which is currently 45% (Dadson et al., 2003).

To explore sediment dynamics within the foreland basin, mass balance calculations were done for a southward propagating orogen model. The total amount of material transported into the basin (according to tectonic scenario of Fig. 12) is compared with the amount of sediment preserved. The theoretical total amount of material, which has been eroded from the orogen since 4 Ma, is estimated by multiplying the integrated area (Fig. 12) with the erosion rate (Table 1). Currently 55% of the annual fluvial sediment discharge is flowing to the west and 45% is drained to the east (Dadson et al., 2003; Liu et al., 2008). Hence the total sediment volume produced by the orogen was corrected for the fluvial discharge flowing to the east.

Figure 12 shows the potential sediment flux into the foredeep coming from the Taiwan mountains as estimated by the model. Computed fluxes vary from 25'000 km3 (for a southward propagation rate of 30 km/Ma and an average erosion rate of 2 mm/yr) and up to 425’000 km3 (for a southward propagation rate of 90 km/Ma and an average erosion rate of 12 mm/yr), although this may be overestimated since it does not take into account the recycling of foredeep sediments. The current river sediment flux during typhoon season (Liu et al., 2008) was taken as an upper boundary for the maximum possible sediment influx, when extrapolated over 4 Ma (i.e., 285'000 km3).

Comparing both, the amount of sediment preserved in the basin (Fig. 4) with the possible amount of sediment carried into the basin (Fig. 12), with a southward propagation rate of 90 km/Ma), and assuming an average upper and lower erosion rate of between 4 to 6 km/Ma (Table 1) since emergence, our calculations suggest that between 25'000 km^3 and 115'000 km^3 of material may have bypassed the foreland basin. If this is correct, it suggests that at least half the sediment eroded from the orogen may not be preserved in the eventual stratigraphic record of the foreland basin. This material is likely longitudinally transported south out of the basin (Nagel et al., 2013), and into the South China Sea. Observations in south-central Taiwan already indicate enhanced southward sediment transport since Late Pliocene marked by increasing amounts of submarine incisions (Fuh et al., 2003; Fuh et al., 1997). The southward sediment transport is also observed in the migration of sediment depocenters and facies belts, mainly driven by the large sediment flux from the orogen (Nagel et al., 2013; Simoes et al., 2007).

4.2. Simulated sediment fluxes

An orogen that produces a steady flux of sediments was modeled for each of the three different growth scenarios in Figure 9 and the volume of material deposited in the basin was calculated for each scenario (Fig. 13). Steady state is established when the elevation of the mountain top reaches a roughly constant value in less than 1 Ma. This is achieved by tuning with the diffusion coefficient for continental sediment transport K, where an increase in K equals an increase in erosion, until a value of K is found that works for all 3 scenarios. Three different models were run, with a mean uplift rate set to 3, 5 and 12 mm/yr. Material is allowed to leave the basin to the south by diffusion. The area of the orogen at each time step is the same for each growth model, thus with identical uplift and erosion parameters the available material at each interval is assumed to be equal. This allows us to compare all three models in terms of only the tectonic growth scenario and longitudinal transport efficiency..

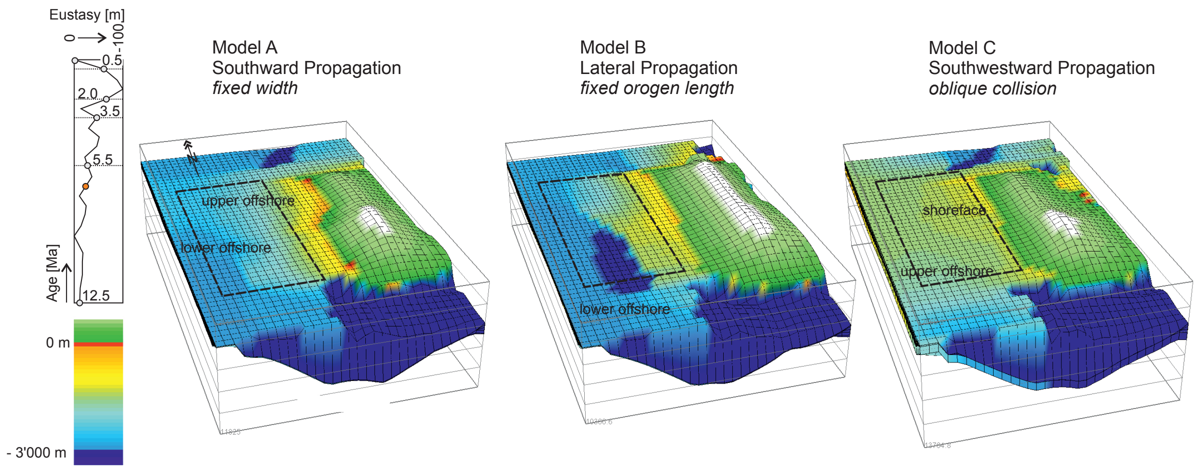

Figure 13: The model setup is shown schematically in Fig. 9. Model A is simulated with a southward propagation rate of 90 km/Ma and a fixed width. In model B the length of the orogen was fixed and only lateral propagation allowed. Model C is a combined " oblique" collision, or southwestward propagation. In all the three models, the final orogen area and final erosional fluxes vary only ba minor amount due to different erosional landscape evolution during relief growth. The area where the simulated foreland basin volume was measured is indicated with a black box.

The sediment volume of the foreland basin produced by each of the three models is shown in Figure 14. The three standard models (southward, lateral, or oblique propagation) tend to overestimate the preserved sediment volume. Southward and oblique propagation achieve a better fit to the observed sediment thickness than lateral propagation. Moreover lateral propagation did not accurately reproduce the foreland basin geometries. The best fit (geometry and volume) is achieved with the oblique collision scenario.

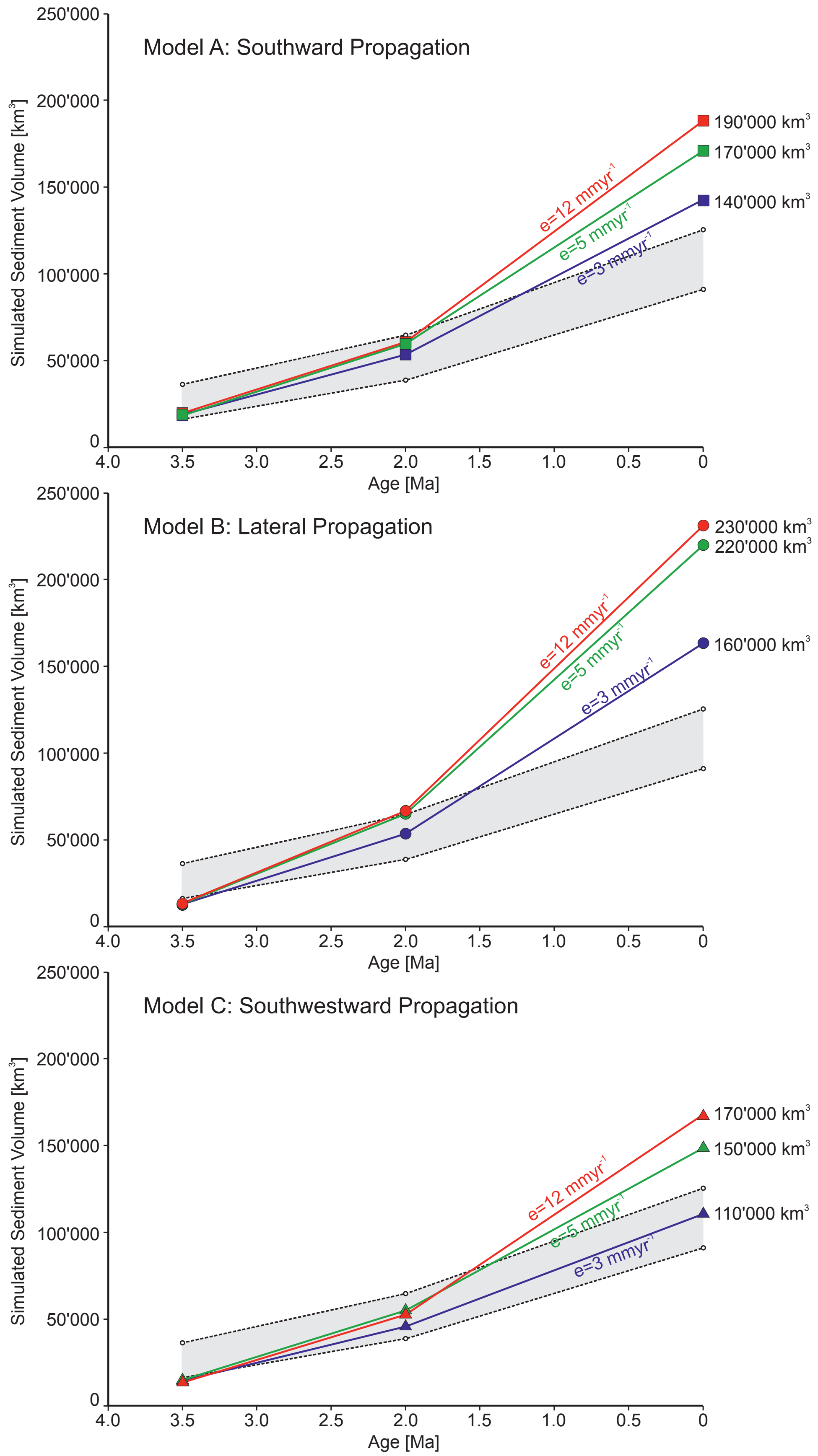

Figure 14: Calculated sediment volume in the foreland basin produced with the three standard collision models (Fig. 9). The preserved sediment thickness in the Taiwan foreland basin is between 70-125'000 km^3 (shaded area, see also Table 3).

The southward propagation models suggest an excess of sediment carried into the basin between 15-80'000 km^3. This amount is in agreement with the theoretical mass balance calculations (Fig. 12). As observed, even though the orogen reached a steady state size as suggested by (Suppe, 1981), due to longitudinal transport the basin never becomes overfilled.

Earlier observations already implied an important longitudinal sediment transport out of the basin and observations from the southwest of Taiwan seem to confirm these predictions (Covey, 1984; Yu and Hong, 2006). Longitudinal sediment transport is common in most foreland basins. A good example is the southern Pyrenees, where longitudinal sediment routing systems dominated a wedge-top depozone, with deep marine sedimentation prevailing (Mutti, 1977, Castelltort et al., 2017). It is important to note in contrast, that an averaged orogen-wide erosion rate of 3 mm/yr produces a sediment volume that is consistent with the preserved sediment volume in the western foreland basin (Fig 14, Model C). This means that, according to our approach, either previous estimates of erosion rate based on thermochronological constraints are too high, or sediment bypass occurred at least for parts of the basin history.

Because of the presence of many submarine canyons draining sediment from the Taiwan Strait to the deeper basin in the Manila trench (Damuth, 1979; Yu and Chang, 2002; Yu et al., 2009), a fundamental unknown is whether one can find there the missing sediment volume arising from our calculations. Sparse literature data are available on the nature of the sedimentary basins in the area of the South China Sea close to Taiwan (Lee et al., 1993; Lin et al., 2008; Yu and Huang, 2009), with a main focus on the Pearl River delta and associated submarine fan deposits (Lüdmann et al., 2001; Su et al., 1989; Xiong et al., 2004) (Li et al., 2008). A topographic map of the submarine regions south of Taiwan indicates a peculiarity in the slope of the South China Sea continental margin compared to its continuation further to the south. This suggests an anomalous accumulation of sediment in this area. Topographic profiles across and along the continental margin (Fig. 15, inset) show that the ocean floor remains at a bathymetry of about -4000 m. As a first order approximation we use the isobath -3600 m and a line roughly parallel to the shelf edge to delimit the contour of this promontory of the continental margin and to compute its volume. The volume enclosed by the area drawn on Figure 15 and using -4000 m as a base elevation represents 28'700 km^3 (15'400 km^3 when -3600 m is used as a base elevation for the calculation).

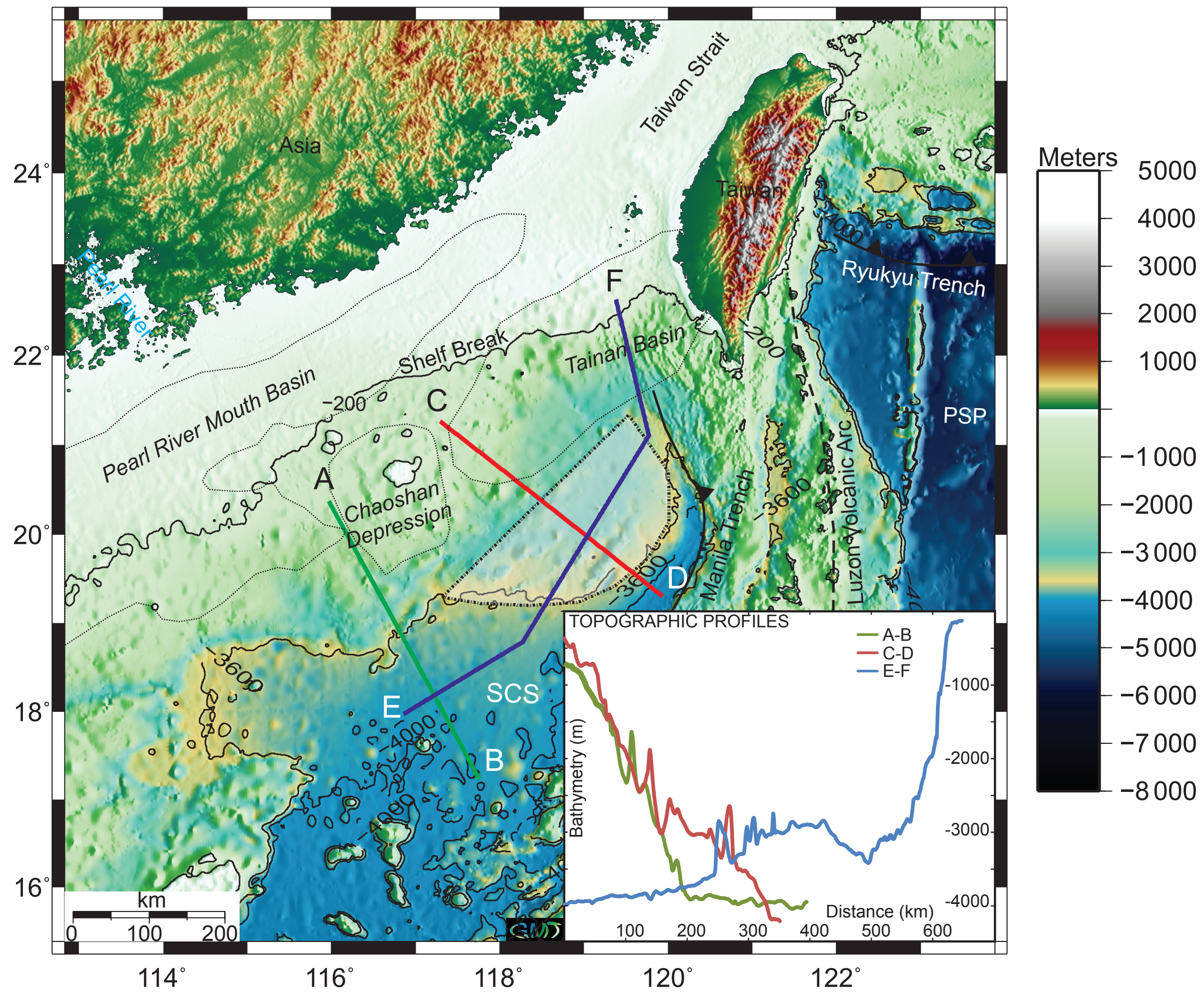

Figure 15: Calculated sediment volume in the northern South China Sea based on topographic profiles across and along the Asian continental margin. The volume stored in the area (dashed line) represents ~ 30'000 km^3.

The volume of this submarine topography is compatible with deposits originated in the Taiwan orogeny that would have bypassed the Taiwan Strait. The outline of the Tainan basin on the topographic map of figure 15 and its southwestward orientation visible on the paleogeographic maps of figure 4 show that the Tainan basin may have constituted a longitudinal through working as a conduit for material sourced in the Taiwan orogenic wedge. In this case, a non-negligible portion of the sedimentary record of mountain building may have been preserved outside of the foreland basin itself. However this hypothesis remains to be tested with future work investigating the sedimentological nature and stratigraphy of this anomalous promontory and look for potential sediment depocenters outside of the Taiwan Strait. This finding outlines the potential complexity of interpreting provenance signals (Romans et al., 2016) in orogen-basin systems with highly dynamic topographic evolution.

5. Conclusions

The sedimentary system of the Taiwan foreland basin is governed by the oblique collision between the Luzon volcanic arc and the Asian passive margin. Different geometrical models of orogen growth and its influence on the basin architecture were tested by means of a stratigraphic modeling approach. We observe that by looking at the sediment volume in the foreland basin and calculating mass flux sediment budgets, a significant (perhaps more than 50%) portion of the sediment eroded from the orogen is not preserved in the stratigraphic record of the immediately adjacent foreland basin. The excess sediment is most likely transported northward into the Okinawa Trough and southward into the South China Sea, where large submarine channel-lobe systems developed. This interpretation is consistent with an increasing amount of submarine incisions since Late Pliocene observed in southwest Taiwan.

We propose that this may be one possible explanation as to why despite reaching a steady state, the basin remains underfilled. We tested three different orogenic growth scenarios with longitudinal transport. While predicted preserved sediment thicknesses exceeded observed sediment thickness, longitudinal transport was efficient enough to keep the basin from overfilling in all three scenarios. However, we find that, despite recent suggestions that collision in Taiwan may have been synchronous along its entire length (Castelltort et al., 2011, Lee et al., 2015), an oblique collision fits better the observed basin architecture.

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